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Bulletin
of
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VOLUME 60 August 1949 NUMBER 8

CONTENTS

Geology of the McCoy area, Eagle and Routt counties, Colorado. By Henry F. Donner 1215–1248

Geology of the Monadnock region of New Hampshire. By Katharine Fowler-Billings 1249–1280

Quaternary geology of the June Lake district, California. By W. C. Putnam 1281–1302

Short Notes

Seismic refraction measurements in the Atlantic Ocean Basin. By Maurice Ewing *et al.* 1303–1304

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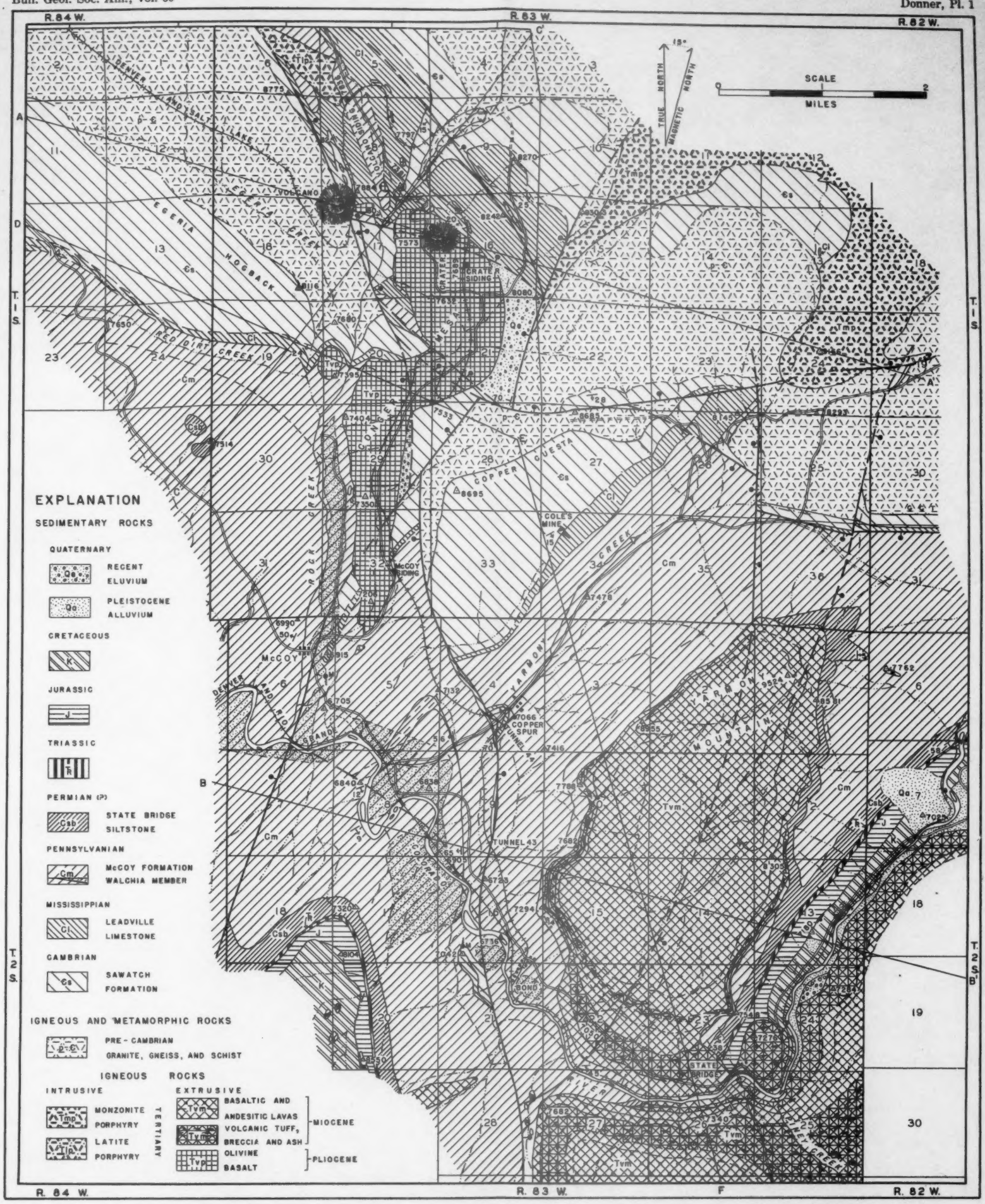

GEOLOGIC MAP OF THE McCOY AREA, COLORADO

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 1215-1248, 5 PLS., 7 FIGS. AUGUST 1949

GEOLOGY OF THE McCOY AREA, EAGLE AND ROUTT COUNTIES, COLORADO

By Henry F. Donner

CONTENTS

	Page
Abstract	1215
Introduction	1216
Location of area and outline of geology	1216
Method of mapping	1216
Previous work and acknowledgments	1217
Topography	1217
Relief	1217
Drainage	1217
Crystalline rocks of Precambrian age	1218
Sedimentary rocks	1218
Cambrian system	1218
Mississippian system	1221
Pennsylvanian system	1222
General considerations	1222
McCoy formation	1223
Pennsylvanian or Permian (?) system	1228
State Bridge siltstone	1228
Triassic system	1229
Jurassic system	1230
Cretaceous system	1230
Tertiary system	1231
Quaternary system	1232
Igneous rocks	1232
Intrusive rocks	1232
Distribution and relation to faults	1232
Microscopic descriptions	1232
Age	1233
Early extrusive rocks	1233
Distribution	1233
Source of flows	1233
Structural relations	1233
Microscopic descriptions	1234
Age	1234
Tilting of lavas	1234
Late extrusive rocks	1235
Distribution and source	1235
Structural relations	1235
Microscopic descriptions	1235
Age	1236
Structure	1237
Pre-Laramide structural setting	1237
Laramide structures	1238
Folds	1238
Faults	1239
Late Tertiary structures	1242
Geologic history and physiography	1242
Economic geology	1244
Copper and lead	1244
Lime	1245
Ballast	1245
Mineral springs	1245
Summary	1246
References cited	1246

ILLUSTRATIONS

Figure	Page
1. Index map showing location of the McCoy area	1216
2. Columnar section for the McCoy area	1220
3. Relation of separate lava flows of Plate 4, figure 1	1236
4. Structure map of the McCoy area	1238
5. Cross sections D-E, E-F, and B-B′ of Plate 1	1239
6. Cross sections C-C′, and A-A′ of Plate 1	1240
7. Suggested movements along fault plane in Plate 5, figure 1	1241

Plate	Facing page
1. Geologic map of the McCoy area	1215
2. General view of the McCoy area and filled limestone cave	1216
3. Photomicrographs of intrusive rocks	1217
4. Structures in Pliocene lava flows	1234
5. Sedimentary rock structures	1235

Abstract

The McCoy area lies in the transition zone between the ancestral Front Range Highland and the Central Colorado Basin. The Precambrian crystallines, here mostly granite, are overlain by sedimentary rocks of Cambrian age. Limestones overlying the Cambrian rocks and previously mapped as Ordovician are assigned a Mississippian age. The Pennsylvanian McCoy formation is re-defined to include over 3500 feet of coarse arkosic sandstones and grits with interbedded shales and limestones. A Walchia bed is recognized as a member of the formation. Red siltstones overlying the McCoy formation conformably are defined as the State Bridge siltstone and are either Pennsylvanian or Permian. Above the siltstone are thin representatives of some of the Mesozoic formations to the west: the Shinarump, Chinle, Entrada, Morrison, Dakota, Benton, and Niobrara formations.

The sedimentary formations were deformed during the Laramide orogeny into open asymmetrical folds and then faulted. The mosaic pattern of the faults is interpreted to mean deformation due to vertical forces rather than horizontal.

A monzonite stock was intruded in Laramide times, after the faulting. Following prolonged erosion, extensive basalt and andesite flows were poured out in miocene time. After another period of erosion volcanic activity broke forth again.

River terraces and three erosion surfaces are recognized. The Black Mountain surface upon which the Miocene lavas were spread was warped before the Sheephorn surface developed. The Blacktail surface was formed some 500 feet below and into this the Colorado River has cut another 500 feet.

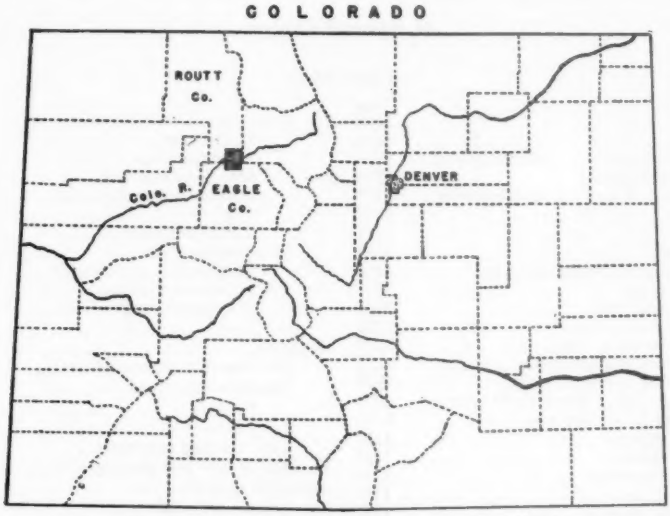

FIGURE 1.—INDEX MAP SHOWING LOCATION OF THE McCOY AREA

INTRODUCTION

Location of area and outline of geology

The area described in this paper includes about 70 square miles near the center of the northwest quarter of Colorado, about 95 miles west of Denver (Figure 1). The Colorado River crosses the southern part; McCoy, Bond, and State Bridge are the largest hamlets.

The area is a hilly, broken, dissected region in an early mature stage of physiographic development, about 12 miles west of the Gore Range, and close to the northwest border of the Southern Rocky Mountain physiographic province. Very little of the region is wooded.

The rocks range from Precambrian to Recent river gravels. The pre-Tertiary rocks are folded and faulted and, in the northern part, intrusion by early Tertiary igneous rocks has caused mild mineralization in some of the sediments. Lava flows of middle Tertiary age cap the highest ground in the southeast. The area sagged locally after their extrusion, with the lowest part along Piney Creek and the State Bridge vicinity. In the northwest, at a much lower elevation and filling an old valley, are late Tertiary lavas. At the north end of these flows and related to them are two volcanic cones.

Method of mapping

The field work in the McCoy area was done during the summer of 1935, with a brief visit to the region in 1936 and again in 1944. For a base map, a lattice of sections was carefully plotted on a plane table sheet using a scale of 2 inches to the mile. Diagrams giving directions and distances of section lines were obtained from the General Land Office, Denver. A number of the section-corner markers were located and flagged in various parts of the area; base

FIGURE 1. VIEW LOOKING SOUTHEASTWARD ACROSS McCOY AREA
Volcanic cones in foreground, Pliocene lavas in valley, Miocene lavas on left skyline.

FIGURE 2. PENNSYLVANIAN RED SEDIMENTS FILLING CAVE IN MISSISSIPPIAN BLUE LIMESTONE, VOLCANO SIDING

GENERAL VIEW OF THE McCOY AREA AND FILLED LIMESTONE CAVE

FIGURE 1. BIOTITE GRANITE, ROCK CREEK VALLEY, ¾ MILE EAST OF VOLCANO; CROSSED NICOLS

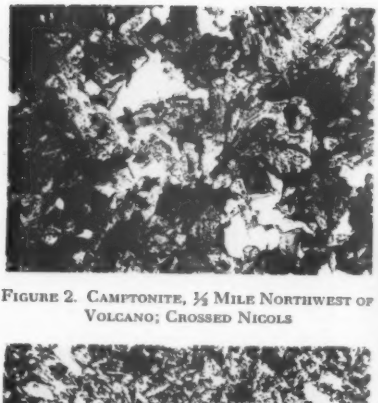

FIGURE 2. CAMPTONITE, ½ MILE NORTHWEST OF VOLCANO; CROSSED NICOLS

FIGURE 3. QUARTZ LATITE PORPHYRY, VOLCANO SIDING; CROSSED NICOLS

FIGURE 4. BASIC HYPERSTHENE ANDESITE, LOWEST FLOW EAST OF STATE BRIDGE; CROSSED NICOLS

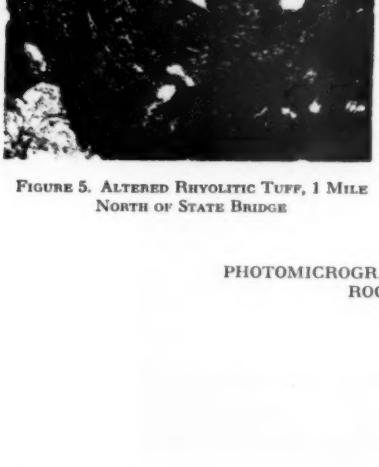

FIGURE 5. ALTERED RHYOLITIC TUFF, 1 MILE NORTH OF STATE BRIDGE

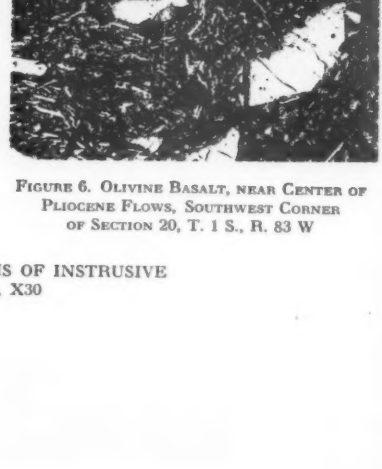

FIGURE 6. OLIVINE BASALT, NEAR CENTER OF PLIOCENE FLOWS, SOUTHWEST CORNER OF SECTION 20, T. 1 S., R. 83 W

PHOTOMICROGRAPHS OF INSTRUSIVE ROCKS, X30

lines were determined from these. Two main base lines were used, one about 3½ and the other about 4½ miles long. A triangulation network was extended over the area by means of a plane-table and telescopic alidade and tied into two United States Geological Survey bench marks in the area to determine elevations. A large number of critical points were located by intersection and resection and their elevations determined. Details were filled in by pacing and using the Brunton compass and aneroid barometer. The Colorado River was plotted on the lattice from Plates 34 and 35 in the United States Geological Survey Water Supply Paper No. 396. Other drainage and roads were filled in from Holy Cross National Forest and Routt National Forest Service maps and other sources. The plotted material was revised when found in error in the field.

Previous work and acknowledgments

Earlier work on the geology of the McCoy area, since the Hayden Survey, consists of reconnaissance studies by Lovering and Johnson (1933, p. 353–374) and by Lovering for the United States Geological Survey. The stratigraphy of the lower part of the Pennsylvanian beds has been studied by Roth (1930, p. 1249–1278), the fauna by Roth and Skinner (1930, p. 332–352), and the beds named the McCoy formation. Since the writer first worked in the area, Bassett (1939), Brill (1942, 1944), and Elias (1944) have worked either in the area or in the vicinity.

The writer is deeply grateful to Dr. T. S. Lovering of the United States Geological Survey for supervision and direction throughout this investigation and for his constructive criticism of the manuscript. To Dr. A. J. Eardley of the University of Michigan thanks are due for his suggestions in the field and his critical reading of the manuscript. The writer wishes to thank Dr. W. F. Hunt of the University of Michigan for suggestions and criticism concerning the igneous rocks and Dr. G. M. Ehlers of the University of Michigan for the identification of the invertebrate fossils. The work of Dr. C. B. Read of the United States Geological Survey, Dr. W. C. Darrah, then of Harvard University, and Dr. M. K. Elias of the Nebraska Geological Survey in the identification of fossil plants is greatly appreciated. To his wife, the writer is grateful for invaluable assistance in the field.

Topography

Relief

The region discussed in this paper is decidedly hilly but not rugged (Pl. 2, fig. 1). The altitude ranges from 6600 feet on the Colorado River about 1 mile southwest of McCoy to about 9600 feet at the top of Yarmony Mountain 3½ miles north of State Bridge, a difference in altitude of 3000 feet. The hills become higher toward the east, and Lone Peak, 12 miles from State Bridge, is on the crest of Gore Range and has an altitude of 11,089 feet. Piney Peak on Piney Ridge, 9 miles east-southeast of State Bridge, is 11,583 feet high. Yarmony Mountain in the southeast is the most prominent elevation within the area. It is separated from the still higher ground bordering the area on the south and east by the Colorado River Valley. The remainder of the area is characterized by hills, hogbacks, valleys, flats, and gorges.

Drainage

The southern part of the region mapped is drained directly by the Colorado River. The river flows southwest to State Bridge where it turns and flows northwest until it leaves the area a mile southwest of McCoy. The central part is drained by Yarmony Creek which flows southwest and enters the Colorado River about 1½ miles southeast of McCoy. The northern and northwestern parts of the area are drained by Rock Creek and its tributaries. Rock Creek flows south from the northern part of the area and enters the Colorado River nearly a mile south of McCoy. Rock Creek in the northwest and Piney Creek in the southeast are the only perennial tributaries of the Colorado within the area.

Only in the northwest does the effect of structure upon drainage become apparent. Egeria and Red Dirt creeks flow parallel to the strike of the beds and Egeria Creek shows a tendency toward trellis drainage.

Crystalline Rocks of Precambrian Age

The Precambrian rocks include granites, gneisses, and schists, and are exposed only in the northern half of the area (Pl. 1) where faulting and folding have brought them above the general level of the surface and erosion has removed the cover of sediments. Excellent exposures occur in Rock Creek and Egeria Creek canyons in the northwest (Pl. 5, fig. 1). Granites and gneisses are by far the most common. One of the most abundant of the foliates found in the McCoy area is a dark-gray medium-grained, gneissoid granodiorite, with about 60 per cent green hornblende. The plagioclase is subhedral and from 2 to 5 mm. in diameter and the anhedral quartz is from 0.5 to 2 mm. Some orthoclase and microcline are present but in amounts less than 5 per cent. The accessory minerals are apatite, zircon, magnetite, and titanite. The petrography of this rock, except for the scarcity of microcline, conforms fairly well with that of the gneissoid Archean quartz monzonite of the Georgetown region described by Ball (Spurr *et al*, 1908, p. 52–54). Rocks believed to be the same age crop out extensively on Boulder Creek west of Boulder, Colorado, and have been named the Boulder Creek granite by Lovering (Lovering and Goddard, 1938, p. 12, 13). Thus the gneissoid granodiorite may be the equivalent of the Boulder Creek granite of the Front Range. This granite gneiss is considered one of the oldest intrusives (Boos and Boos, 1934, p. 306) in the Idaho Springs formation which is the oldest Precambrian formation in Colorado.

A pink, medium-grained to coarse-grained granite consisting largely of pink microcline and quartz with scattered specks of biotite is common in the area studied. This granite weathers more readily than some of the other types of Precambrian rocks and where the contact between Precambrian and Cambrian rocks reaches the creek level in Rock Creek valley, about three-fourths of a mile east of Volcano, it is strongly weathered to a depth of 5 feet below the contact, suggesting that Cambrian weathering was progressing more rapidly than erosion in that particular spot when burial occurred.

Microcline, ranging from 0.5 to 5 mm. across, constitutes about 50 per cent of the fresh granite; quartz, ranging from 0.5 to 3 mm., forms about 30 per cent; oligoclase, 1 to 2 mm. in size, constitutes about 10 per cent; and biotite about 3 per cent of the rock (Pl. 3, fig. 1). The accessory minerals consist of apatite, magnetite, titanite, and hematite. Hematite dust is profusely scattered throughout the rock. The petrography conforms well with that of the Pikes Peak granite, named and described by Mathews (1895, p. 472; 1900, p. 217–225) in the Pikes Peak region, also described by Ball (Spurr *et al*, 1908, p. 57–58) as Rosalie granite in the Georgetown quadrangle and by Lovering (1935, p. 12–13) in the Montezuma quadrangle. In northern Colorado and Wyoming a similar granite is called the Sherman granite. According to Boos (1934, p. 306–307) the Pikes Peak granite intruded the Boulder Creek granite-gneiss and thus is younger.

A few dark, fine-grained camptonite dikes from 6 to 12 inches wide cut the pink granite locally; these dikes are about 40 per cent green hornblende, 30 per cent is sericitized feldspar, chiefly oligoclase, and 20 per cent quartz (Pl. 3, fig. 2).

An elongated knob about one-half mile northwest of Volcano is composed of a pink, fine-grained, granitic rock of sugary texture, probably an aplitic facies of the pink granite. Pegmatite dikes were found in the granite exposed on the hillside 2 miles south of Crater. These were composed entirely of white quartz, books of muscovite measuring 4 inches across, and some biotite.

Schist is present in the McCoy area only as local patches and streaks included in the granite.

No detailed work was done on the Precambrian rocks; they were mapped and treated as a single unit.

Sedimentary Rocks

Cambrian system

The Cambrian system is represented by the Sawatch formation in central Colorado. It was so named by Eldridge (Emmons *et al*, 1894, p. 6) because of its persistent occurrence around the flanks of the Sawatch Range. Its age was set as Upper Cambrian by Cross (1894, p. 1–2). In Manitou Park and near Manitou Springs, brachiopod shells, *Lingulepis* and *Obolella*, were

found and in Garden Park a trilobite, *Ptychoparia*. These forms, according to Cross, indicate Upper Cambrian deposits.

Behre (1932, p. 58) states that a small collection of brachiopods from this formation in the Leadville district indicates Upper Cambrian age but does not give the genera. He separates the Sawatch formation into a quartzitic lower member and an upper member which he designates as the Peerless shale member.

Both members of the Sawatch formation were recognized by Lovering and Johnson (1933, p 359) in the McCoy area north of McCoy. They state that the formation thins out and disappears north and east beyond the limits of the area covered in this paper and thickens to about 350 feet near Glenwood Springs, 45 miles southwest of McCoy. The author attempted to map the quartzite member and Peerless shale member as separate units, but they are so nearly identical in the area covered that it is not practical to separate them. The Sawatch formation was, therefore, mapped as a single unit.

The Cambrian rocks are mainly confined to the northern half of the area. They are exposed in the extreme north, along the crests of Egeria Hogback and Copper Cuesta, and on the face of some of the faults in the central and northern half of the area. The formation rests unconformably upon a comparatively flat Precambrian surface.

The total thickness of the Sawatch formation is about 270 feet. It is light-gray to white and largely composed of quartz conglomerate with interbedded sandstone. The top 110 feet also contain interbedded micaceous shales. The bedding is thin to massive and is locally crossbedded. The finer sandstones are somewhat micaceous. Pebbles of quartz, averaging the size of peas and scattered in a finer-grained groundmass, are common, especially near the base. Near the center are interbedded bands of friable sandstone and conglomerate from one-half to 1 foot thick, which appear to be poorly cemented. These sandstone bands weather out first and leave grooves in the vertical exposures. Locally the formation is a dense white quartzite, but in many places the rock breaks around the pebbles rather than through them and does not have the appearance of a quartzite. Near the top are bands of conglomerate with quartz pebbles as large as 2 inches in a finer-grained groundmass.

Section of Sawatch formation in Rock Creek valley 300 yards northeast of road bridge east of Volcano

Disconformable contact with Leadville limestone

Units	Thickness Feet
Sawatch formation	
9. Conglomerate, light gray to white, quartz. Interbedded shales and sandstones. White mica is abundant in the shales. Conglomerate and sandstone largely altered to quartzite. Bands of conglomerate a few inches thick contain quartz pebbles as large as 2 inches scattered in a finer groundmass	25
8. Conglomerate and sandstone, light gray to white, quartz, largely altered to quartzite. Thin interbedded, gray, micaceous shales	57
7. Similar to division 8 except not as much shale and no marked alteration to quartzite	25
6. Conglomerate and sandstone, thin to massively bedded light gray, quartz. Interbedded layers, ½–1 foot thick, of both conglomerate and sandstone, very friable, crumbling in the hand. Cement has apparently been altered and these bands weather out first, leaving grooves in the exposures. In general rock breaks around quartz pebbles instead of through them	25
5. Conglomerate and sandstone, thin to massively bedded light gray, quartz, crossbedded locally and breaking around pebbles instead of through them, in general. Finer sandstones are somewhat micaceous. Conglomerates contain quartz pebbles averaging the size of peas scattered in a finer groundmass	133
4. Conglomerate consisting of 2-inch subangular quartz pebbles in a finer quartzite matrix	0.5
3. Arkosic rock, light gray, somewhat kaolinized. Distribution is local, in places as much as 1 foot thick	0.5
Total	266

COLUMNAR SECTION FOR McCOY AREA

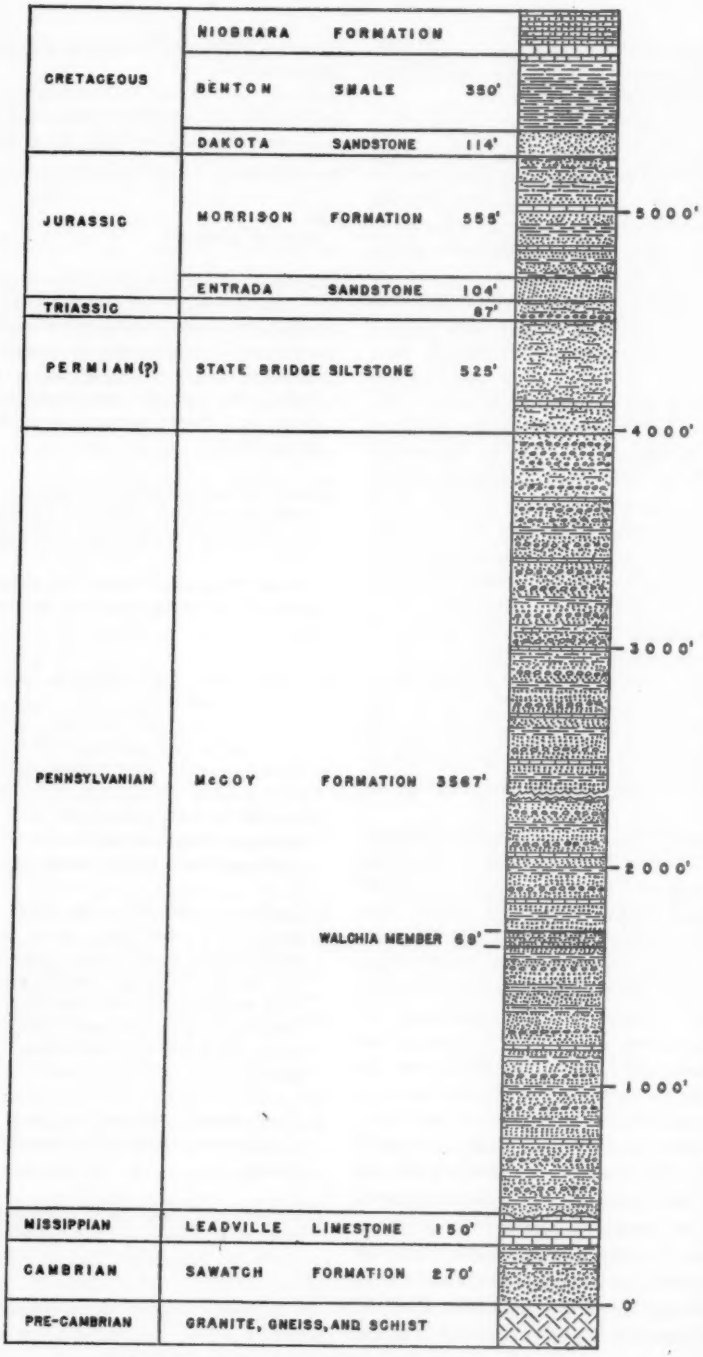

FIGURE 2.—COLUMNAR SECTION FOR THE McCOY AREA

Unconformable contact with Precambrian rocks
Precambrian rocks

Units	Thickness Feet
2. Granite, weathered, easily crumbling under the hammer	5
1. Granite, pink, medium-grained, with veins and patches of mica schist	—

Mississippian system

The lower part of the back slopes of Egeria Hogback and Copper Cuesta are composed of a bluish-gray limestone. This rock is also exposed in the north-central part of the area and along the edges of some of the main faults. It lies disconformably upon the Sawatch formation of Upper Cambrian age and is here considered to be the Leadville limestone of Mississippian age. A section of the limestone, measured about 300 yards north of the bridge across Rock Creek east of Volcano, is about 80 feet thick. A section in the railroad cut southwest of tunnel 44 at Copper Spur is 152 feet thick. A quarter of a mile east of McCoy only erosional remnants of the Leadville remain, and part of the exposed Sawatch formation is overlain by Pennsylvanian rocks. North and east of the area the Leadville disappears within a few miles. Westward, in the vicinity of Dotsero, Bassett (1939, p. 1860) measured a section over 190 feet thick consisting of dolomite and limestone.

The upper surface of the Leadville limestone is very uneven and, therefore, its thickness varies greatly. Surface and underground channels and caves were developed before the Pennsylvanian sediments were deposited. Excellent sections of caves in the limestone, filled with Pennsylvanian red shales and mudstone, are exposed in railroad cuts between Crater Siding and Volcano Siding. About 300 yards northeast from the cone of Crater is an example of a cave which was probably only partly filled with Pennsylvanian red sediments and later collapsed due to pressure from the sides. The sediments within are contorted and the limestone above is arched. About 450 yards north of Volcano Siding another fine example of such caves or underground channels is exposed (Pl. 2, fig. 2).

The Leadville limestone is a massive to thin-bedded, fine-grained, bluish-gray limestone. Locally it is lithographic. In places it is light gray and in exposures east and northeast of McCoy the weathered rock is yellowish brown. About 300 yards downstream from the bridge across Rock Creek east of Volcano, the lower 20 feet of the Leadville consist of interbedded shales and limestones. Some of the limestones are coarse-grained and contain fossils. Black chert nodules are common near the top of the limestone as on the hillside northwest of Copper Spur. Calcite veins are also of frequent occurrence. The Leadville limestone shows no special topographic expression in the area studied although it appears to be more resistant than the overlying Pennsylvanian beds.

Section of Leadville limestone one-half mile northeast of Volcano

Disconformable contact with McCoy formation

Units	Thickness Feet
4. Limestone, thin to massively bedded, fine-grained to lithographic, bluish to brownish gray. Bottom few feet lined with calcite crystals. Interval partly covered toward top	60
3. Limestone, thin-bedded, very fine-grained, bluish gray	2
2. Shale, chocolate red. This bed varies from a thin film to 6 inches and is filled with angular pebbles of limestone up to an inch across where it reaches its maximum thickness	0.5
1. Limestone, very fine-grained, bluish to brownish gray. Massive at base and thin-bedded near top with few thin beds of shale. Lower part contains cavities, 3–4 inches across, lined with calcite crystals and contains a few thin veins of calcite	19
Total	81.5

Disconformable contact with Sawatch formation

Section of Leadville limestone west of tunnel 44 at Copper Spur

Units	Thickness Feet
Disconformable contact with McCoy formation	
5. Limestone, fine-grained, dark gray	60
4. Fault gouge	15

Units	Thickness Feet
3. Limestone, fine-grained, light gray	53
2. Fault gouge, green	4
1. Limestone, fine-grained, light gray to light yellow, changing to limy sandstone with quartz pebbles near base	20
Total	152

Disconformable contact with Sawatch formation

Lovering and Johnson (1933, p. 359, 363) considered the blue limestones overlying the Sawatch formation in the McCoy area to be Ordovician. They are practically unfossiliferous and thus age determination is very difficult. The writer was fortunate in finding a small exposure in Rock Creek valley half a mile east of Volcano, where interbedded shales and coarse limestones form the lower 20 feet of the formation. In a coarse limestone, about 10 feet above the contact with the Sawatch formation, fossil fragments were found. Also a poorly preserved specimen of a brachiopod was found which Dr. G. M. Ehlers identified as a long-hinge, multiplicate species of *Spirifer*. The limestone, therefore, can not be Ordovician. The writer is inclined to consider it of Mississippian age because (Kirk, 1931, p. 237) "It should be borne in mind that the Devonian when present north of an east-west line lying somewhere north of Ouray is usually dolomitic, whereas the Mississippian on an average is low in magnesian content." At Leadville, the type locality for the Leadville limestone, the rock is a dolomite. The black chert nodules near the top of the limestone in the McCoy area are characteristic. They have been noted by Emmons (1886, p. 63, 66), by Brainerd and Johnson (1934, p. 535–537), by Emmons, Irving, and Loughlin (1927, p. 33, 34), and others.

Fossil caves in the Madison limestone, filled with material of the Fountain formation above, have been reported by Brainerd, Baldwin, and Keyte (1933, p. 391, 392) in the Beulah district. Brainerd and Johnson (1934, p. 540), in referring to the Mississippian sediments of Colorado, state:

"Locally the old limestone surface was eroded into sink holes and a semi-karst topography. Good examples of solution effects can be observed in the mines at Gilman and in the Aspen district where the overlying Pennsylvanian shales have filled sink holes and other irregularities in the old surface."

This feature is also observed in the McCoy area. The color of weathered outcrops is also characteristic of the Mississippian limestone. The yellowish-brown exposures east and northeast of McCoy may be an exception. Kirk (1931, p. 226, 239) says that in weathered outcrops the Devonian tends to weather brownish or yellowish, whereas the Mississippian has a decidedly bluish cast; the brownish-weathering rocks near McCoy are possibly remnants of the Devonian limestone. They have been mapped tentatively as Leadville limestone although they could not be traced horizontally into the blue-weathering type.

The name Leadville limestone is used in this study, in the restricted sense given by Kirk (1931, p. 239), to designate the Mississippian limestone which has the blue-gray color characteristic of the Leadville. According to Brainerd and Johnson (1934, p. 541) the paleontologic evidence so far obtained shows the Leadville limestone to be of early Mississippian age, equivalent to the Madison of Wyoming, probably equivalent to the Kinderhook, and possibly to the lower Burlington of the Mississippi Valley. Faunal lists are given by Girty (1904, p. 217–231) and Emmons (1886, p. 66).

Pennsylvanian system

General considerations.—Originally the author divided the great thickness of sedimentary rocks lying between the Leadville limestone and the basal conglomerate of the Triassic into two divisions. The lower 2340 feet were called the Weber (?) formation and the upper 1750 feet the Maroon formation because of stratigraphic position and lithologic characters. This also was in harmony with the names used for these formations in other parts of central Colorado. The division between the two formations was arbitrarily set at the base of a coarse conglomerate containing many large pebbles of limestone. This conglomerate seems a little more persistent than the others in the series. Above this division black shales cease and, except the top 525 feet which are siltstones, the overlying rocks are mainly coarse sandstone and conglomerate with only a few thin beds of shale and limestone.

Two mappable units of the Maroon formation were recognized in the McCoy area. The lower member was named the Rock Creek conglomerate because of its fine exposures along the west bank of Rock Creek north of McCoy, and the upper member was named the State Bridge siltstone because of its excellent exposure on the hillside behind State Bridge Hotel.

After more extensive work in the field and discussions during the years of 1936 to 1940 with T. S. Lovering, K. G. Brill, Jr., and C. F. Bassett who had worked or were working in the same general vicinity, it was decided to raise the State Bridge siltstone member to the rank of formation and to extend the McCoy formation as named and defined by Roth (1930, p. 1265; Roth and Skinner, 1930, p. 332) to include all the beds between the Leadville limestone and the State Bridge formation. This gives a total thickness of at least 3567 feet instead of 1000 feet (Roth and Skinner, 1930, p. 332) or 1011 feet (Roth, 1930, p. 1266, 1267) for the McCoy formation. Such a division is much more practical in the field as there is a distinct lithologic change in passing from the coarse arkosic grits of the upper McCoy formation into the State Bridge siltstone. The stratigraphic names Weber (?), Maroon, and Rock Creek were therefore abandoned. Further work by Brill (1944, p. 635–638) after these discussions has shown that bringing the State Bridge member up to formational rank was justified. He finds a wide lateral distribution of these rocks and a thickness of over 6000 feet at Woody Creek about 40 miles south of State Bridge.

McCoy formation.—The lower two-thirds of the McCoy formation is widely distributed in the central and southern part of the area, along the southern boundary of Egeria Hogback, and in patches in the northern part. The upper third is exposed in the east-central, west-central, and southwest part of the area. The formation rests disconformably upon the Leadville limestone which has a very irregular surface. North of the area it disappears within a few miles, and 6 miles east of Yarmony Mountain it overlaps the underlying formation, resting on Precambrian rocks.

A complete section could not be found in the McCoy area because of the numerous faults. The nature of the exposures and the lack of persistent beds made the measurement of the thickness of the upper McCoy conglomerates rather difficult, and an indirect method had to be used for part of it. An approximate thickness of at least 3567 feet for the entire McCoy formation was obtained.

The formation is composed of arkosic sandstones and conglomerates, gray to salmon to maroon colored, with interbedded fossiliferous, dark-gray limestones, and gray, black, and maroon, micaceous shales. The conglomerates and sandstones are abundantly cross-bedded on a large scale. The character of the beds varies greatly in short distances both vertically and horizontally. The formation is in part apparently of piedmont origin. The last several hundred feet at the top consist almost entirely of pink to red, coarse sandstone and arkosic grits and conglomerates highly cross-bedded with very little shale and limestone and no fossils. This upper series is fairly resistant and often forms cliffs and canyons.

In the railroad cut north of the highway crossing at Bond is an example of a local unconformity in the McCoy. The lower beds are practically horizontal and the upper ones dip to the north, as seen in the section, at an angle of about 15 degrees. A valley on each side of the section makes it impossible to trace the beds laterally to see what has taken place. Another example may be seen about 100 yards southeast of tunnel 43 on the Denver and Salt Lake railroad line. These are not delta structures since the beds are cut off at the contact with those below. It is believed that these local unconformities are diagenetic, due possibly to channeling of the soft sediments and creep of large blocks or masses into these channels, the bottom creeping faster than the top. An example of this in recent soil and slope wash may be seen in the southwest corner of section 21, T.1 S., R.83 W. A railroad cut, 25 to 30 feet deep, has been made into this material, which is now creeping out into the cut. Large cracks, considerable distances from the edge of the cut, are forming on the top.

A persistent horizon occurs at about 1222 feet above the base of the McCoy formation. Measurements of its thickness show a variation from 68 to 117 feet. The bottom 68 feet are very persistent; some of the sandstones in it show peculiar raised markings that look like forms

squeezed out of a cake decorator and range from very small ones to some an inch across and several inches long. The horizon consists of alternate layers of dark, greenish-gray shales only a few millimeters thick and light-gray micaceous sandstone ranging from a fraction of an inch to 2 inches thick and containing fossil plants (Pl. 5, fig. 2). Thicker beds of shale separate the above units at intervals. This horizon was so persistent, characteristic, and conveniently distributed that the base of it was used as a key horizon in plotting the structure contour map (Fig. 4). It is here named the Walchia bed because it contains fossil remains of the plant *Walchia;* it is considered a member of the McCoy formation.

Section of McCoy formation

Measured in sections 8 and 17 of T. 2 S., R.83 W.

Conformable contact with State Bridge siltstone

Units	Thickness Feet
94. Sandstone, arkosic grits, and conglomerates, pink to red, strongly cross-bedded, with few interbedded shales and limestones in lower portion	640
Measured on west side of Rock Creek bridge at McCoy	
93. Limestone, light gray, dense, no fossils. (Leadvilleoides of Ehlers)	4
92. Conglomerate and sandstone, mostly pink to maroon, arkosic, cross-bedded, with little shale	184
91. Limestone, dark gray, not persistent, barren	1
90. Conglomerate, pink to gray to maroon, arkosic; cross-bedded sandstone and micaceous shale	30
89. Limestone, dark gray, fine-grained, with fossils. Lens shaped and splits into 2 laterally	3
88. Similar to unit 90	241
87. Shale and sandstone, red and green, micaceous. Interbedded nodular limestone containing fossils. Coarse sandstone at top also contains fossils probably obtained from limestone conditions just below	24
86. Conglomerate, grits, and sandstone, maroon to gray, arkosic; interbedded red shale	85
85. Conglomerate, coarse persistent with large limestone pebbles	16
84. Similar to unit 86	54
83. Limestone, dark gray, very impure, nodular	0.8
82. Conglomerate, cross-bedded, gray, arkosic	3
81. Shale, green with some interbedded nodular limestone. (Few fossils?). (Roth and Skinner No. 178?)	7
80. Conglomerate, grit, sandstone and shale, maroon to gray, arkosic	58
79. Shale and sandstone, brownish maroon to gray with micro-fossils	39
78. Limestone, dark gray, nodular appearpearing, with interbedded shale. Fossiliferous. (Probably Roth and Skinner No. 177)	6
77. Grits, sandstone, and shale, maroon and gray, arkosic	113
76. Sandstone, light gray, medium to coarse, arkosic, rather soft at top	3.5
75. Shale, maroon and green, micaceous	1
74. Sandstone, blue gray, calcareous medium-grained, weathering brown	0.7
73. Shale, dark gray to greenish gray, micaceous. Largely covered	47
72. Limestone and interbedded shale, dark gray, fossiliferous. (Probably Roth and Skinner No. 175)	9
71. Sandstone, coarse, gray, arkosic	15

Units	Thickness Feet
70. Shale and limestone nodules, dark gray. Mostly covered and variable in thickness. Fossiliferous. (Probably Roth and Skinner No. 174)	5
69. Sandstone and shale, gray, coarse. (Cross over from north to south side of road)	7
68. Shale, black to dark gray, micaceous. Interbedded, light gray, fine sandstone.	40
67. Sandstone, gray, coarse, arkosic	8
66. Shale and sandstone, black to gray, interbedded. Mostly covered	15
Measured 1 mile northwest of Bond	
A break of unknown thickness probably exists between unit 66 and unit 65	
65. Shale, red and gray sandstone	28
64. Shale, interbedded, green; dark, massive, fossiliferous limestone	10
63. Shale, dark to brick red, micaceous; massive gray sandstone; pink grits striped with maroon and cross-bedded. (At 70 feet are 3 limestone beds, 1–3 inches thick, interbedded in shale)	140
62. Limestone, dark, massive with interbedded shale, fossiliferous	6
61. Shale, green	3
60. Shale, dark red	8
59. Grits, pink, arkosic, maroon streaks	33
58. Shale, maroon, interbedded with gray micaceous sandstone	36
57. Shale, black, interbedded with light-gray micaceous sandstone	57
56. Grits and conglomerates, pink, arkosic, cross-bedded	100
55. Shale, black, fissile; interbedded micaceous, light gray sandstone	60 ⅔
54. Shale, green and black, micaceous, containing fossil plants	28
53. Grits and conglomerates, pink, arkosic.	32
52. Sandstone, grits and conglomerates, light gray, grading into dark shale at top	21
51. Shale, carbonaceous	25
50. Grits and conglomerates, pink, arkosic.	30
49. Sandstone, thin, micaceous, light gray. Interbedded greenish-gray shale. Sandstones contain fossil plants and many peculiar markings. (Walchia Bed)	68
48. Conglomerate, cross-bedded, grayish red, arkosic with pebbles as large as 7 inches across	13
47. Shale, maroon and green, micaceous. Thin-bedded sandstone and conglomerate near center about 2–4 feet thick	24
46. Similar to unit 48	8
45. Shale, black, micaceous; fine sandstone.	8
44. Similar to unit 48	10
43. Shale, maroon, micaceous	6
42. Conglomerate, light gray to maroon, highly cross-bedded and containing thin layers of dark-maroon shale	13
41. Shale, blue and maroon, micaceous	2.5
40. Limestone, interbedded with shale similar to unit 41. Contains fossils	4.5
39. Conglomerate, bluish gray, arkosic	1.5
38. Shale, dark gray, micaceous	1.5
37. Similar to unit 39	5
36. Shale, greenish gray and maroon, micaceous	19
35. Conglomerate, cross-bedded, maroon and gray, arkosic; interbedded maroon shale	40

Units	Thickness Feet
34. Limestone, dark gray, nodular, sandy. Reworked material with fossil impressions	1
33. Sandstone, gray and red, micaceous. Numerous fucoid casts at base	8
32. Shale, greenish gray and maroon. Nodular limestone concretions and layers in shale appear to have been reworked	8
31. Conglomerate, highly cross-bedded, light gray and maroon, arkosic; some interbedded shale	19
30. Shale, maroon, micaceous; interbedded light gray and maroon arkosic conglomerate	33
29. Conglomerate, cross-bedded, pink, maroon, and light gray, arkosic; little interbedded shale and sandstone	62
28. Shale and sandstone, dark gray, micaceous	12
27. Shale, maroon, micaceous with ½ foot of conglomerate near center	23
26. Shale, maroon, gray sandstone and conglomerate	45
25. Sandstone, light gray, many wormlike casts. Weathers into discs	5
24. Shale, green, micaceous	20
23. Conglomerate, yellowish pink, cross-bedded, many large pebbles	21
22. Shale, sandstone, and conglomerate, green to dark gray	40
21. Limestone, dark gray, fine-grained. Contains *Jonesina*. (Probably Roth and Skinner No. 184)	1
20. Shale, green to dark gray	11
19. Sandstone, yellowish gray; conglomerate with some interbedded shale	9
18. Shale, green to dark gray, micaceous	6
17. Similar to unit 19	11
16. Similar to unit 18	5
15. Conglomerate and sandstone, arkosic, maroon and light gray	28
14. Shale, green to dark gray, micaceous	31
13. Limestone, brown-weathering. *Myalina* bed on top. (Probably Roth and Skinner No. 183)	1
12. Similar to unit 14	2
11. Sandstone, thin-bedded, light gray, micaceous	19
10. Shale, dark, micaceous	9
9. Conglomerate and sandstone, light gray; little interbedded shale	22
8. Shale, dark, micaceous; some interbedded sandstone and with 1–3 layers of calcareous beds which are very irregular in lateral distribution and contain fossils in places. (Mostly covered.)	55
7. Conglomerate, gray, very coarse near base	4
6. Shale, sandstone, and conglomerate, maroon and gray. (Largely covered.)	235
5. Sandstone and conglomerate, gray, resistant; interbedded maroon shale. Note: To complete the section a quartzitic conglomerate bed was traced north 1 mile to the southern part of section 4, T.2 S., R.83 W. where the remainder of the McCoy is exposed in a railway cut.	195
4. Conglomerate, quartzitic, reddish yellow, very resistant	24
3. Sandstone, shale and fault gouge, dark red	11
2. Sandstone and grit, light gray, coarse	15
1. Sandstone and shale, dark red and yellow mottled	75
Disconformable contact with Leadville limestone	
Total	3567

As noted in the above section, a break of unknown thickness probably exists between unit 65 and unit 66. Brill made a correlation of fossiliferous, purple limestones containing red chert by making a study of the dolocasts of these limestones in this area (written communication, 1939). If his correlation is correct the break is only about 15 feet. The writer is inclined to believe, however, that it is much greater than this, otherwise the "Walchia Bed" should be exposed on the west side of Rock Creek near where it enters the Colorado River.

Because of the folded nature of the sediments in the McCoy area and the alternation of shales, coarse sandstones, and conglomerates in the McCoy, the surface is very irregular where this formation is exposed. The shales weather out first and leave the resistant rocks as ridges where they are bowed up in folds. On the west side of Yarmony Mountain where the McCoy is protected by the lava sheet capping the mountain, it appears as steep, bare hillsides.

Roth (1930, p. 1265–1267) describes a section at McCoy and named it the McCoy formation. In his published section, which is 1011 feet thick, he includes a partial faunal list and from it concludes that the section seems to be of lower Cherokee age. Roth and Skinner (1930, p. 332–336) list 55 genera and 73 species of megafauna and microfauna from the McCoy formation and give the localities where the collections were made. Among other correlations they state that the lower Fountain in Colorado may be correlated with the McCoy formation. The stratigraphically highest locality in which Roth collected is at McCoy and lies about 58 feet below what the writer considered to be the dividing line between the Weber (?) formation and the overlying Maroon formation in his early work before these names were dropped. It is at least 2274 feet above the Leadville limestone or what Roth terms Cambro-Ordovician undifferentiated. The McCoy fault, a south-southwest trending fault with downthrown side on the west passes through the east side of McCoy. It may explain the difference between Roth's 920–930 feet and the writer's 2274 feet for the thickness of the same series of beds. The stratigraphically lowest collection made by Roth, 600 feet east of tunnel No. 43 on the Denver and Salt Lake Railway, is about 670 feet above the Leadville limestone or base of the McCoy. The Copper Spur fault, a north-south fault with downthrown side on the east, passes between tunnel 43 and the locality where the collections were made and cuts out 600–700 feet of strata. This fault again may explain the discrepancies between Roth's findings and those of the author; Roth places his lowest collection about 525 feet above the base of the McCoy formation and the highest only 400 feet above the lowest (Roth and Skinner, 1930, p. 335), whereas it should be at least 1600 feet between the two.

Fossil plants collected in the key horizon, 1222 feet above the base of the McCoy, were sent to C. B. Read of the U. S. Geological Survey, Washington, D. C. and to W. C. Darrah, then of Harvard University for identification. Read tentatively regards the well-preserved fragment of a fern as *Sphenopteris decipiens* Lesquereux and states that the other plant fragments on the slab are not identifiable. One might be a *Walchia*, but it could also be a small leafy twig of *Lycopodites*. The evidence at hand, he states, is not very weighty for an age assignment, but he is inclined to regard the flora as older Pennsylvanian, and either lower or middle Pottsville. Darrah tentatively identified the specimens as a *Callipteris* group, *C. naumanni* or *C. lyratifolia* and *Walchia piniformis* and believes the flora are Permian because of the presence of *Callipteris*. This plant horizon lies about 1050 feet stratigraphically below the highest faunal collection made by Roth and 560 feet above his lowest faunal collection and, according to his conclusions, should therefore be of Cherokee or Pottsville age.

Moore, Elias, and Newell (1936, p. 1–31) report Permian flora from the Pennsylvanian rocks near Garnett, east-central Kansas. The rocks are considered to be Middle Pennsylvanian (upper part of Lansing group of Missouri series) and among the flora listed are: *Taeniopteris, Neuropteris, Sphenopteris, Walchia piniformis*, but no *Callipteris*. The Permian-Pennsylvanian boundary has been defined in recent years on the basis of the first appearance of the fern *Callipteris conferta*. The authors interpret these peculiarities of the flora as due to environment factors. The flora is undeniably of Permian type but lived in Pennsylvanian time. Elias (p. 12) states:

"The predominance of *Walchia piniformis*, with addition of other conifer-like remains and of *Taeniopteris*, gives the flora so striking a "Permian" aspect that if a paleobotanist were asked to determine the age without knowledge of where these plants came from, he would almost certainly conclude that the assemblage belonged to Permian time."

The same specimen containing the plant remains studied by Read and Darrah was sent to M. K. Elias in 1937. He distinguished the following forms on the specimen: *Walchia piniformis*, *Odontopteris* of *O. affinis-O. schlotheimi* group (also comparable to *Sphenopteris germanica* Weiss), *Cardiocarpus* cf. *orbicularis*. He would not put much weight on this small flora as evidence for geologic age of the rocks. Working in the McCoy area more recently Elias found *Walchia* flora near Bond and in an abstract (1944, p. 1468) fixes their age as late Cherokee. A forthcoming paper by him will throw more light on the flora and fauna of this region.

The lower part of the McCoy formation, approximately the 2340 feet originally mapped as the Weber (?) formation by the writer, may be the equivalent of the Weber (?) formation in the Salt Creek area, Mosquito Range, as determined by Gould (1935, p. 976–977). This part of the McCoy, however, is much coarser and about 600 feet thicker. It was evidently deposited under more variable and shallower water conditions, as indicated by its cross-bedded, coarse sandstones and arkosic conglomerates which are missing in the Salt Creek area. Limestone forms a very small portion of the formation in the McCoy area but a large portion in the Salt Creek area. The lower and middle part of the Weber (?) formation of the Mosquito Range is ragarded as Pennsylvanian and equivalent to the middle Pottsville and part of the upper Pottsville of the East by Johnson (1934, p. 33) who gives a full faunal list (p. 31, 33).

The Coffman conglomerate member of the Maroon formation of Gould's section in the Salt Creek area is lithologically similar to the upper McCoy formation. This member is apparently more grayish and thinner than the dark-red to maroon to brick-red rocks of what the author originally called the Rock Creek conglomerate in the McCoy area. The upper McCoy formation may be the equivalent of the Coffman conglomerate. Plant impressions from the lower part of the Coffman have been identified by David White as *Calamites* and *Lepidodendron* and tentatively dated as "Pennsylvanian, probably not lower Pennsylvanian nor yet too near the top of the Pennsylvanian" (Gould, 1935, p. 985). These correlations of the lower and upper McCoy formation with formations in the Salt Creek area are in general agreement with those of Brill (1942, 1944) working in the Gore area.

Brill (1944, p. 626, 627, 639) states that the Belden shale formation in the Gore area probably grades into the lower McCoy formation and that its fauna is indicative of Des Moines age. He also states (p. 631, 640) that the Maroon formation overlying the Belden shale is equivalent to the upper McCoy formation. He concludes that although the upper Maroon is without fossils it is probably all of Des Moines age. In the light of the fossil evidence and the correlations mentioned above the writer believes the McCoy formation is of early Pennsylvanian age, probably all of Des Moines age.

Pennsylvanian or Permian (?) system

State Bridge siltstone.—This formation is so named because of its excellent exposure on the hillside behind State Bridge Hotel. It is exposed along the east side of Yarmony Mountain, in the southwest, and in small erosional remnants in the west-central part of the area. It rests comformably upon the underlying McCoy formation. At State Bridge the siltstone is 525 feet thick, and 5 miles northeast in section 4, T.2 S., R.82 W. it is 517 feet thick. It is composed mainly of thin-bedded, micaceous, brick-red siltstone, shale, and fine, limy, sandstone. In the top portion some beds contain excellent oscillation ripple marks. Approximately 140 feet from the base is an impure dense limestone which at State Bridge is 7 feet thick. It appears to be more resistant than the surrounding shales and siltstones, as it forms prominent ledges. Five miles north-east it is only 1–2 feet thick. This bed is significant because it is the only one in which fossils have been found. Locally, as at State Bridge, the brick-red siltstone is bleached to a purplish-gray or a yellow-

ish-gray. This change is also seen in smaller areas along the east slope of Yarmony Mountain, but in other exposures the brick-red color is persistent.

Section of State Bridge siltstone one-fourth mile northwest of State Bridge

Disconformable contact with Triassic

Units	Thickness Feet
6. Shale, brick red and siltstone, limy, abundant mica	86
5. Siltstone and shale, purplish gray, limy, beautiful ripple marks on some beds. Gray shales with blotches of deep purple	53
4. Shales, yellow and gray, thin bedded; siltstones, limy	88
3. Shale, brick red; siltstone, brick red, limy, micaceous. Interbedded with small amount of gray shale and siltstone	150
2. Limestone, impure, resistant, dense	7
1. Similar to unit 3	141
Total	525

Conformable contact with McCoy formation

The State Bridge siltstone may be equivalent to either the Chubb siltstone or Pony Spring siltstone as named by Gould in the Salt Creek area. The age of the Chubb siltstone is tentatively considered to be Permian by Gould (1935, p. 988) on the basis of impressions of coniferous plants, identified by him as *Walchia* (?). The Pony Spring siltstone is placed in the Permian because of the presence of *Walchia piniformis* and *Walchia gracilis*, which according to David White, who identified the fossils, "are not known from the Pennsylvanian anywhere. *Walchia gracilis* is found in abundance in some localities in the south-west Permian" (Gould, 1935, p. 995). Other fossils from this member have been examined by David White and are said by him to resemble *Ulmannia* and *Voltzia* and are tentatively placed in the Permian.

The presence of Permian flora in the Pennsylvanian rocks of Kansas (Moore *et al*, 1936), with an abundance of *Walchia piniformis*, and the presence of this plant with other Permian plants in the McCoy formation below beds containing marine fauna of Des Moines age weaken considerably the evidence that the Chubb and Pony Spring siltstone members of the Maroon formation are Permian. The writer is of the opinion that the State Bridge siltstone could be Pennsylvanian. Permian is, therefore, followed with a question mark when referring to the age of the State Bridge siltstone.

Brill (1942, p. 1393; 1944, p. 636) found internal and external molds of the pelecypods *Myalina* sp., *Pleurophorus* sp., and *Aviculopecten* sp. in the impure dense limestone bed 140 feet above the base of the State Bridge siltstone, unit 2 of the type section of this formation. These are the only fossils known in the State Bridge formation in the McCoy area. Newell, who examined the fossils, stated, in a written communication to Brill, that they might be middle or late Pennsylvanian or Permian. Brill (1944, p. 636, 640) believes that the indications are that the State Bridge formation is equivalent to the Phosphoria formation of northwestern Colorado and would therefore be of Permian age. This probably is as near to an age determination as can be obtained until diagnostic fossils are found in the State Bridge formation.

Triassic system

Rocks of Triassic age were first recognized by Lovering and Johnson (1933, p. 369) at State Bridge in the McCoy area. They are exposed in a narrow strip from State Bridge northeastward between the Colorado River and Yarmony Mountain and in the southwest part of the area. At State Bridge there are 90 feet including a basal conglomerate 10–15 feet thick. This mottled pink and white conglomerate is composed of quartz and chert pebbles as large as 1 inch. Fragments of fossil wood 10 inches in diameter were found in it. The conglomerate is missing locally. The beds above it are a purplish-red mudstone and shale. Above these is a dark-red mudstone which reaches a maximum thickness of 10 feet in the southern end of the exposure and thins toward the northeast until it disappears within a few hundred yards.

Five miles northeast of State Bridge, in the southwest corner of section 4, T.2 S., R.82 W., the Triassic is only 40 feet thick and consists almost entirely of the basal conglomerate with some coarse sandstone but very little mudstone or shale. Still further north the formation disappears.

According to Reeside (Lovering and Johnson, 1933, p. 369), these Triassic mud shales are probably of Chinle age and the basal conglomerate is of Shinarump aspect. The Triassic beds rest disconformably upon the underlying State Bridge formation as evidenced by the basal conglomerate.

Jurassic system

In the McCoy area the Jurassic system is represented by the Entrada sandstone and the Morrison formation. These formations were also recognized at State Bridge and vicinity by Lovering and Johnson (1933, p. 369).

The Entrada sandstone has about the same distribution as the Triassic beds and rests upon them with a suggestion of a slight angular unconformity. The exposure northwest of State Bridge Hotel shows the Triassic beds thinning toward the north. About 7 miles northeast of State Bridge the Entrada overlaps the Triassic and rests upon the State Bridge siltstone. Still further northeast it is overlapped by the Morrison formation. It is of fairly uniform thickness, measuring about 104 feet which is its thickness at State Bridge. The Entrada is a soft, medium to coarse, salmon-colored sandstone strongly cross-bedded. It is easily distinguished from the beds above and below by its characteristic salmon color.

The distribution of the Morrison formation is the same as that of the Triassic and Entrada. No break is apparent between the Entrada and the overlying Morrison. It overlaps the Entrada and Triassic several miles northeast of the area and rests on the State Bridge siltstone. Only the lower 216 feet of the Morrison are exposed in the hill behind the State Bridge Hotel, the remainder is covered with eluvium and lava. The best place to measure the total thickness of the formation was found outside of the area, 5 miles northeast of State Bridge in the southwest corner of section 4, T.2 S., R. 82 W. Here it is 555 feet thick. The Morrison consists of interbedded gray, limy sandstones; gray limestones; and variegated, maroon, chocolate, and greenish-gray, nonmicaceous shales. The beds often show beautiful ripple marks and are well exposed in the gulch northwest of State Bridge Hotel.

Partial section of Morrison formation 300 yards northwest of State Bridge Hotel

Units	Thickness Feet
6. Lava and eluvium	—
5. Sandstone, white to pink, soft, limy	13
4. Clay shale, variegated maroon and greenish gray	22
3. Limestone, gray, separated by a few inches of irregularly bedded shale at intervals of 1–3 feet	15
2. Shale, maroon, chocolate and gray, nonmicaceous, 2–5 feet thick. Alternating with limestone, light gray, sandy in places, 1–5 feet thick	52
1. Sandstone, light-colored, limy, very soft toward bottom, interbedded with maroon shale in top portion and greenish gray in bottom portion with very little shale in the center. Many of the beds show excellent ripple marks	114
Total	216

Conformable (?) with Entrada sandstone

The top 339 feet of the Morrison are lithologically similar to that portion of the formation given in the above section. Its shaly character causes it to weather into smooth slopes, and it is rather difficult to get a complete uncovered section. Fragments of vertebrate fossils are present in the Morrison formation at State Bridge. The formation is lithologically similar to that at Morrison, Colorado, the type locality.

Cretaceous system

The Cretaceous system is represented in the McCoy area by the Dakota quartzite, the Benton shale, and the Niobrara formation.

The Pierre shale and higher beds are exposed a short distance east of the area and also a few miles south. Only the lower 175 feet of the Niobrara are present within the area mapped or in the immediate vicinity.

Along the northwest side of the Colorado River, extending from near State Bridge northeast, the Dakota quartzite forms the crest of a ridge nearly parallel to the river. The beds here are vertical or dip at very high angles; the quartzite is very resistant, giving rise in places to almost knife-edge ridges. In the southwestern part of the region the beds are nearly horizontal and form a ledge on the hillside. The rock is almost entirely a massive light-gray quartzite. Locally it contains a little interbedded shale. It is about 110 feet thick and is very persistent. It is regarded as the Dakota quartzite because of its position between the Morrison below and the Benton shale above and its lithologic similarity to the Dakota of the Aspen district (Spurr, 1898, p. 41) and other localities in central Colorado. No unconformity is apparent between the Morrison and the Dakota in exposures examined within the area.

From about a mile northeast of State Bridge continuing northeast until the edge of the area is reached, the Colorado River valley is cut into Benton shale for the most part. The strip between the river and the ridge of Dakota quartzite is Benton shale covered, for the most part, with eluvium or slope-wash. Exposures are scarce since the shale weathers easily and soon becomes covered. This formation is also present above the ledge of Dakota in the southwest part of the area where it forms smooth gradual slopes. The Benton shale rests conformably upon the Dakota quartzite and is about 350 feet thick. It consists almost entirely of dark-gray to black fissile calcareous shales. A dark-brown fetid limestone marks the top. Its stratigraphic position and lithologic similarity to the Benton shale of the Montezuma district (Lovering 1935, p. 18–20) and of the Aspen district (Spurr, 1898, p. 41) and other localities in central Colorado justify its being called the Benton shale in the McCoy area.

A small amount of Niobrara is exposed in the southwest part of the area; other outcrops are found along the river northeast of State Bridge and mostly on the southeast bank. The thickest exposure, 175 feet, is about 1¼ miles northeast of State Bridge on the east bank of the river. The contact with the underlying Benton shale is just at the water's edge. It is a conformable contact. As exposed in the area mapped, there is a basal gray limestone about 25 feet thick, followed by 100 feet of light-gray, thin-bedded limestone, sandy limestone and dark-gray limy shale. The next 50 feet are more calcareous and thicker bedded, and the remainder of the formation is absent or covered. The name Niobrara formation is used for these rocks because of their position above the Benton shale and their similarity to occurrences of Niobrara rocks in other parts of Colorado.

Tertiary system

No sedimentary deposits of the Tertiary system older than Miocene were found in the McCoy area. High up on the hill, about 800 feet above the river and one-half mile northwest of State Bridge, where a horizontal lava sheet covers the turned-up, truncated sedimentary beds, patches of coarse waterworn gravel lie between the sedimentary rocks and the lava. These gravels are several feet thick in places but are very irregular in distribution. They and the overlying lavas are probably of Miocene age. It is believed that in this particular area erosion took place during most of Eocene and Oligocene time, and any Tertiary material that was deposited was eroded before Miocene time. The locality at State Bridge is the only place where these gravels were observed. Interbedded between some of the separate laval flows are gravel beds and thick beds of volcanic ash, breccia, and tuff which, south of the area, are very sandy and clayey and are light yellow to gray. They may also be Miocene deposits.

A gulch, cut through the southern end of the late lava flows half a mile northeast of McCoy, exposes the contact of the lava with the underlying material which is here unstratified, dark-red gravel, sand, and silt with boulders 2–3 feet in diameter consisting mostly of quartzite. This contact is about 400 feet above the Colorado River and the lavas are considered to be of Pliocene age. The deposits upon which they rest are also probably Plio-

cene. Evidence for ages of the lavas will be found in the discussion of the extrusive rocks.

Quaternary system

The Pleistocene deposits in the McCoy region are found along the Colorado River and Rock Creek valleys in the form of terraces. Terraces occur at two different elevations, the higher one is of destructional origin whereas the lower is constructional and contains the Pleistocene deposits. The formation and age of these terraces and the pediments will be discussed in the section on geologic history and physiography.

An excellent exposure of these deposits may be seen near the center of section 7, T.2 S., R.82 W. in a railroad cut along the west side of the river. The exposure is about 300 feet long and nearly 100 feet high. The horizontally bedded gravels rest on Niobrara beds dipping southeast at an angle of 70 degrees. The contact is about 25 feet above the present river level, and the top of the terrace is about 200 feet above the water. The terrace in which this section is cut is one of the best in the area. It is about half a mile wide at the river and slopes up to the mountain for three-fourths of a mile at an angle of 5 degrees. Across the river from it is a similar one sloping in the opposite direction at an angle of 5 degrees. On the west side, about a fourth of a mile back from the river, Morrison sandstone, dipping at a very high angle, projects several feet above the level of the terrace, forming monadnock-like elevations above the terrace surface.

The deposits of this terrace, as seen in the railroad cut, consist of unassorted fine gravel, pebbles, and boulders up to 3 feet across. Many kinds of rock are represented, such as quartzite, schist, granite, lava, etc. Above the lowest 6 feet, pockets of fine sand and silt are scattered here and there. Still higher up the beds become better stratified and much more of the finer material is present. Near the top of the exposure yellowish-green clay beds alternate with gravel beds. The edge of the terrace is exposed in this cut and shows a veneer of recent red slope-wash from 2 to 10 feet thick covering the gravel. Farther down the river some of the hillsides show yellowish areas in the ordinary red slope-wash which are due to the Pleistocene yellow silt exposed where the veneer of recent material has washed away. Several exposures of silt and gravel may be seen along the road from Yarmony Siding to State Bridge. The early deposits were coarse gravel and the later ones silt and finer material, interbedded with gravel.

These gravels do not always occur as terraces. Three-fourths of a mile northeast of State Bridge, on the high and steep west bank of the river, the gravels occur as a shell on the face of the bank, resting against lava and older sediments which formed the bank just before the gravels were deposited.

Good exposures may be seen at Bond, at the northwest corner of section 16 and the center of section 8 of T.2 S., R.83 W., also just northeast of McCoy and in places along the east bank of Rock Creek.

The most recent deposits consist largely of river gravels and red earth, pebbles, and boulders which cover the rocks, and the gravel where it has not been removed by erosion.

Igneous Rocks

Intrusive rocks

Distribution and relation to faults.—The intrusive rocks are found in the northeast and north part of the area. A monzonite porphyry stock, most of which is outside the region mapped, has two apophyses which enter the area and appear to be closely related to the faults of the region. The one near the north-central part of the area (Pl. 1.) is approximately parallel to the northern extension of the McCoy fault which forms its western boundary for a considerable distance. The eastern one appears to cover the Yarmony Park fault, in part, and has a smaller apophyse that extends along the fault, indicating that the intrusion is younger than this fault. The Yarmony Park fault apparently affects the McCoy fault, and must therefore be younger. The McCoy fault must, therefore, have preceded the intrusion also.

Microscopic description.—The stock-like intrusion in the northeast consists of a light-colored porphyritic rock with a very fine crystalline groundmass. The abundant light-pink phenocrysts of andesine reach 5 millimeters and the much less common phenocrysts of dark hornblende are noticeably smaller. The gray

groundmass of microcrystalline orthoclase constitutes about 50 per cent of the rock. No quartz was found. The accessory minerals are magnetite, apatite, titanite, and zircon. The rock is evidently a monzonite porphyry. All the high peaks in the northeast, both inside and outside of the area, are composed of this resistant porphyry. In the eastern projection, inclusions of hornblendite, from an inch to a foot, are scattered through parts of the rock.

Immediately northwest of Volcano Siding and trending in a northwesterly direction is a short, thick, dike-like intrusion about a mile long, terminated at its northwest end by a fault. It is a light-gray to light-blue, fine-grained slightly porphyritic rock with phenocrysts of andesine, biotite, and quartz ranging from 0.5 to 2 mm. (Pl. 3, fig. 3). The microcrystalline groundmass, which constitutes about 80 per cent of the rock, is composed almost entirely of orthoclase, with many tiny shreds of biotite, and probably some quartz. The accessory minerals are apatite and garnet. The rock is a quartz latite porphyry and is probably a more silicic phase of the magma from which the monzonite porphyry originated.

Age.—The youngest formation cut by the monzonite porphyry in the area mapped is the Leadville limestone of Mississippian age. The quartz latite porphyry cuts the McCoy formation of early Pennsylvanian age. The relation of the intrusives to the faults indicate that they were either contemporaneous with or later than the faults. The faulting took place after the folding. (*See* under Laramide structures.) The folding involved rocks at least as late as the Niobrara formation. The intrusives are therefore not earlier than late Cretaceous or early Tertiary age. This is in agreement with the age assigned to the belt of intrusive porphyries and monzonites extending from Boulder, Colorado southwest to Leadville and west by Lovering (1935, p. 26), by Crawford (1924, p. 377–378), by Spurr, Garrey, and Ball (1908, p. 67–71, 109–110, 128–130) and others.

Early extrusive rocks

Distribution.—Basaltic and andesitic lavas with interbedded rhyolitic and trachytic tuffs, breccia, and ash occupy the southeastern part of the area mapped. They are only a part of a much larger mass lying to the southeast, which is approximately 15 miles long and 10 miles wide. The Colorado River has cut through the lava, separating the part that forms the top of Yarmony Mountain from the main mass, except at State Bridge east and west of which the river bed is on lava. There are at least nine separate lava flows. A persistent bed of tuff and breccia lies between the fifth and sixth flows from the bottom. Their thicknesses range from 25 to 150 feet.

Source of flows.—A stream has cut into the lava beds and exposed what appears to be a small volcanic neck about 1½ miles east-southeast of State Bridge, just outside of the area mapped. This exposure is a few hundred yards east of John M. Regnold's house in the southwest corner of section 30, T.2 S., R.82 W. Beds of lapilli and volcanic ash are exposed on the northwest side of the projecting mass which is oval in cross-section. The beds dip toward the center at an angle of about 45°. The top contains many volcanic bombs of various sizes. This may have been a source of some of the ejectamenta found in the region.

It is believed that the chief source of the lava was farther southeast because the main mass is in that direction and the early flows thin out and disappear as they are traced northwest to State Bridge. A peculiar exposure of wrinkled lava in the northeast corner of Eagle County may represent the disturbed material at or near a vent that supplied some of the lava flows. It is in the northern half of section 9, T.2 S., R.81 W., about 10 miles east-northeast of State Bridge. These are the only suggestions of sources of the volcanic material that were found.

Structural relations.—At State Bridge the lower surface of the lava reaches the river bed: a half mile north, this surface is 800 feet above the river. Northeastward, the contact gradually rises. One mile west of State Bridge the contact with the underlying sediments is again exposed above the river level and rises gradually along both banks as it is followed westward. Southeast of State Bridge, along Piney Creek, only lava is exposed. The pre-lava drainage was evidently different from that of the present course of the Colorado River.

The first flow to enter the area was very narrow, filling only the pre-lava stream or river channel. Each succeeding flow overlapped the

one below and spread further because it was higher up in the valley. Not until the eighth flow, counting the tuff bed, did the lava reach the extreme north end of Yarmony Mountain, 4 miles north of State Bridge. These upper flows have been eroded far back from the Colorado River, so the whole sequence is not present in any locality (Fig. 5; Pl. 1).

Microscopic descriptions.—The volcanic material consists of basaltic and andesitic lavas, with interbedded trachytic and rhyolitic tuffs, breccia, and ash. Hypersthene andesite is probably the most abundant rock. Commonly it is porphyritic and slightly vesicular. The fine, crystalline groundmass constitutes from 75 to 85 per cent of the rock and is composed of plagioclase laths commonly from 0.05 to .2 mm. long, augite granules and prisms, magnetite, and hypersthene ranging from 0.1 to 0.3 mm. A small amount of interstitial glass is also present. Plagioclase near oligoclase in composition makes up about 55 per cent, augite 20 per cent, and hypersthene about 3 per cent of the groundmass. The phenocrysts are zoned andesine from 0.5 to 3 mm. in size, pseudomorphs of iddingsite after olivine, 0.1 to 2 mm., and small crystals of hypersthene (Pl. 3, fig. 4). Accessory minerals are magnetite and apatite needles. Secondary calcite, hematite, and serpentinitic-chloritic alteration products are common.

The basalts are commonly medium fine-grained, somewhat porphyritic, gray-black rocks. The groundmass has a felted texture and is composed of labradorite laths, averaging about a millimeter long, and irregular grains of augite, usually from 0.1 to 0.2 mm. in size. Magnetite in grains about 0.1 mm. in size is moderately abundant and brown glass is common in small patches generally constituting less than 5 per cent of the rock. Labradorite makes up about 60 per cent, augite about 15 per cent, and phenocrysts of olivine, largely altered to iddingsite and ranging from 0.5 to 2 mm., about 10 per cent of the rock. Secondary products are chloritic-serpentinitic material. The rock is a coarse-grained olivine basalt.

The top of the persistent tuff and ash bed which covers the fifth lava flow 1 mile southeast of Copper Spur, may be taken as typical of the volcanic ash. It has a rough, irregular fracture and dull, earthy appearance, the matrix can be scratched with a knife and is light gray with a pink tint. The inclusions consist of crystal fragments, chiefly sanidine, rock fragments, and sub-angular lava inclusions ranging from a fraction of a millimeter to an inch across. There is almost no quartz. The yellowish-brown glassy groundmass has numerous shards containing dark borders. It also contains small fragments of orthoclase, oligoclase, andesine, microperthite, magnetite, hematite, biotite, and zircon. Vugs and crevices in the rock are lined and partly filled with vein quartz, chalcedony and opal.

Locally this strongly altered, partly silicified, trachyte tuff contains masses of silicified rhyolitic tuff from 50 to 75 feet high and about the same distance across. They are well exposed at Copper Spur and in the ash bed on the east side of Yarmony Mountain about one mile north of State Bridge. They vary from red to gray. The broken surface is smooth and glassy in appearance with semi-conchoidal fracture and cannot be scratched with a knife. About 40 per cent of the rock consists of rock and crystal fragments, 25 per cent of which are sanidine ranging from 0.1 to 2 mm. in size and usually corroded. Free quartz is abundant. Both andesine and oligoclase, ranging from 0.1 to 2 mm., and some orthoclase are present. Irregular fragments of fine-grained basalt, from 2 to 5 mm., biotite, magnetite, hematite, garnet, zircon, and apatite are other constituents present. Cavities lined with chalcedony and opal were filled with quartz. (Pl. 3, fig. 5)

Age.—Wilson (1939, p. 315) assigns a Lower Miocene age to the tuff beds on the basis of a new species of dog *Cynodesmus casei* found in the tuff about 5 miles south of State Bridge. At least 3600 feet of sediments were eroded from parts of the area after the Laramide revolution, and the somewhat incised surface was formed. According to Lovering (1929, p. 108; 1934, p. 23), volcanic activity and pronounced uplift occurred in Miocene time in the Front Range. The lavas and tuffs forming the cap of Yarmony Mountain were probably extruded at this time and deposited in an essentially horizontal position.

Tilting of lavas.—The uplift that followed the volcanic activity in the Front Range appears to

FIGURE 1. ISOLATED MASS OF LAVA SHOWING PECULIAR STRUCTURE, LOOKING NORTH

FIGURE 2. BUCKLING OF LAVA BAND NEAR CONTACT WITH SEDIMENTS ON RIGHT
South of Crater, looking east.

STRUCTURES IN PLIOCENE LAVA FLOWS

FIGURE 1. FAULT IN PRECAMBRIAN-CAMBRIAN ROCKS ¼ MILE NORTH OF CRATER

FIGURE 2. STEEPLY DIPPING WALCHIA BEDS OF MCCOY FORMATION ON WEST SIDE OF COPPER SPUR FAULT
Northwest corner of section 16, T. 2 S., R. 83 W.

SEDIMENTARY ROCK STRUCTURES

have been more of a warping in the area mapped. The zone passing along the Colorado River from near McCoy in a southeast direction to south of State Bridge and along Piney Creek is the zone of greatest sag. The lavas northeast of this zone dip to the southwest at an angle of about 9 degrees. They have probably sagged in the zone and been elevated northeast of it. The lavas southwest of this zone dip more gently and to the northeast. The lava flows therefore appear to form a broad east-west trending syncline with the lowest part southeast of State Bridge just outside the area of this report.

The tilted lava may be seen best across the river east of State Bridge where the surface is terminated at the top by Piney Ridge, the highest ground in the vicinity, and at the bottom by Piney Creek. The bending along the zone of sag and the crumpling that would result may have been influential in determining the course of Piney Creek and the right-angle turn of the Colorado River near State Bridge.

Late extrusive rocks

Distribution and source.—Extending from a point between Crater and Volcano to about half a mile east of McCoy is a mass of lava roughly the shape of a question mark. It is about 4 miles long and nearly a mile wide in the north, narrowing gradually to a point at the southern end. In the north it is about 250 feet thick; at the southern end it is only about 15 feet thick. The surface is fairly even and slopes very gradually to the south. It contains the best ranches in the area and is called Conger Mesa (Pl. 2, fig. 1). At the northern end are two volcanic cinder cones, one, resting on the lava, called Crater, and the other, about 1 mile west-northwest of it, called Volcano. The base of Volcano is about 300 feet higher than that of Crater, and the cone is somewhat larger. The scoria is being quarried from its eastern side and used for ballast on the railroad line. It is believed that both centers contributed lava to the flow which followed the course of the old drainage system southward. This drainage was east of the present drainage, and in the north was about 200 feet above Rock Creek, in the south about 300 feet above it. Rock Creek cut its valley along the western edge of the lava to a depth of 450–500 feet below the lava plain forming a cliff on the west side of the plain. The area between Crater and Volcano was undoubtedly completely covered with lava at one time, but now only two remnants occur because of the erosional activity of Rock Creek.

Structural relations.—Just north of the junction of Rock and Egeria creeks, Rock Creek has cut through and isolated a small mass of lava which shows a very peculiar structure (Fig. 3; Pl. 4, fig. 1).

It has been conjectured that the first lava flow came down the valley from the north and temporarily filled the valley to the top of the isolated remnant. When the volcanoes ceased pouring out lava, the level of the flow lowered in the middle of the stream as it continued to flow. Along the edges and in protected nooks such as the locality of the isolated remnant, a plastic crust "a" formed on the surface. As the lava drained out below, the crust sank to the bottom to form the band "a". The second flow came down the valley and the supply of the lava was halted. A plastic crust "b" was formed on the surface which sank in the middle as the molten lava below drained out down stream. A similar sequence of events accompanied successive bursts of lava until the sixth or final flow, which cooled in place to form the crust "f" and the mass "F" below it. The ridge to the west of the remnant kept the level fairly constant. If the lava rose above this, or too much was squeezed out by overlying flows, it overflowed onto the other side of the ridge. Plastic bands of lava showing columnar structure existed in the main lava mass and figure 2 of Plate 4 shows such a band considerably wrinkled up where it was pushed against the sediments to the right.

Microscopic descriptions.—The lavas which make up these late flows have produced fine-grained rocks, gray black near the top, black in the center, or that portion which does not have columnar jointing, and blue gray near the bottom. They are somewhat porous, and are quite vesicular at the top. Caves measuring several feet across are common in exposed faces; some may be original but many are due to weathering. Olivine phenocrysts up to 5 mm. may be seen in hand specimens.

The gray-black lava near the top of the flows has a porphyritic texture and many vesicles from 0.1 to 0.5 mm. in diameter. The elongated ones reach a length of 1 millimeter.

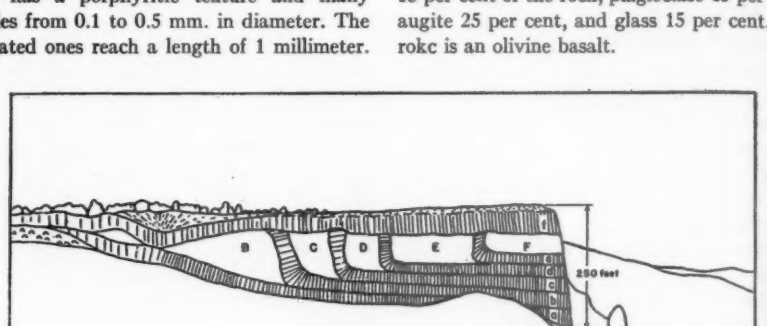

FIGURE 3.—RELATION OF SEPARATE LAVA FLOWS OF PLATE 4, FIGURE 1

The parts of the lava labeled with capital letters are assumed to be parts of the flows labeled with the same lower case letters. The bands a, b, c, d, e, and f, with good columnar structure, are lighter in color and break when hit by a hammer much more easily than the black, irregularly jointed portions B, C, D, E, and F.

The groundmass, constituting about 90 per cent of the rock, is microcrystalline and consists of feldspar laths with a maximum length of 0.1 mm., augite grains, and numerous fine grains of magnetite. Hematite and possibly some glass are present. About 40 per cent of the rock is feldspar, 30 per cent augite, 5 per cent magnetite, 12 percent vesicles and 10 per cent olivine subhedral and euhedral phenocrysts ranging from 0.1 to 2 mm. The rock is an olivine basalt.

The black lava near the center of the mass, where it is harder and does not have columnar jointing (similar to B, C, D, E, F, in Fig. 3), has a porphyritic texture (Pl. 3, fig. 6). The groundmass is very fine-grained and has a felted texture consisting of laths of plagioclase with maximum length of 0.2 mm., and laths and microlites of augite. These laths show a flow structure. Magnetite is present as fine grains and grains averaging 0.1 mm. There is considerable brown glass present. Phenocrysts of subhedral olivine from 0.1 to 2 mm. and an occasional augite phenocryst make up about 15 per cent of the rock, plagioclase 45 per cent, augite 25 per cent, and glass 15 per cent. The rokc is an olivine basalt.

Near the base of the mass, the blue-gray lava with columnar jointing has a porphyritic texture with a very fine-grained holocrystalline groundmass consisting of laths of plagioclase about 0.2 mm. long, and grains, laths, and microlites of augite. The laths show a flow structure. Grains of magnetite average about 0.2 mm. in size. Phenocrysts of subhedral olivine from 0.1 to 2 mm., and a few phenocrysts of augite 1 mm. in size, make up about 15 per cent of the rock, plagioclase 45 per cent, augite 35 per cent, and magnetite 5 per cent. The rock is also an olivine basalt.

The cones of Crater and Volcano are made up almost entirely of scoria, some black and some red. Bombs of all sizes and shapes may be found, ranging from the size of nuts to 3 feet in diameter. A small pile of scoria was found in the southeast corner of section 16, T.1 S., R.83 W. and may possibly represent a very small eruption as it is favorably situated in regard to faults.

Age.—The early lavas of the McCoy area

were tilted and deeply dissected. In one of the valleys thus formed the late lavas were poured out. Unlike the older lavas, which are widespread and for the most part cap the higher ground such as Yarmony Mountain, the late lavas are inlaid, although not in contact with the older lavas. It is clear, therefore, that the inlaid lavas are appreciably younger than the capping lavas.

The cinder cones Crater and Volcano look surprisingly fresh and undissected, and by comparison with cones of similar aspect they would be judged Pleistocene, if not late Pleistocene. The lava flows down the valley from these cones, however, appear older. A good soil has formed on the flows; there are no remaining pressure ridges, spatter cones, lava blisters or other surface forms of basic lava flows; Rock Creek in its lower 3 miles has dissected the flows 450–500 feet deep, and this must have taken considerable time. The age of this episode of dissection will only be determined by a study of the terraces of the Colorado River and by relating them to the glacial deposits of nearby ranges. At the present the flows look like early Pleistocene or late Pliocene, but the cinder cones seem to be almost Recent.

Lovering (1929, p. 104) has noted a marked uplift in the Front Range to the east in late Pliocene or early Pleistocene time, with consequent dissection. If this uplift affected the McCoy area, as seems possible, then the late flows pre-date the uplift and would be late Pliocene.

Structure

Pre-Laramide structural setting

According to Melton (1925), Lovering (1929), Ver Wiebe (1930), Heaton (1933), and others, the area occupied by the present Front Range of Colorado, including the Park and Gore ranges, was high ground throughout Paleozoic and most of Mesozoic time. This high area is known in the literature both as the Front Range Highland and as the Colorado Highland. Its position coincided approximately with that of the present Front Range. During late Pennsylvanian and early Permian time, the uplift reached its maximum height and from then on to Middle Cretaceous time was gradually overlapped by the Permian and Mesozoic sediments. It was finally buried by the Dakota sandstone, and Upper Cretaceous sediments continued to accumulate over its site. During late Cretaceous and early Tertiary time the Laramide revolution built the present Front Range through lateral compression (Chamberlain, 1919; 1925), or vertical uplift (Lee, 1923), or a combination of the two. This was accompanied by considerable folding and faulting followed by extrusive and intrusive igneous activity.

Southwest of the Front Range Highland is the Central Colorado Basin and beyond that the Uncompahgre Highland in the southwest of Colorado. The Colorado basin was one of deposition throughout Paleozoic and Mesozoic time. The boundary between the Front Range Highland and the basin shifted back and forth during Paleozoic and Mesozoic time (Lovering and Johnson, 1933).

The McCoy area lies in the transition zone between the Front Range Highland and the Central Colorado Basin, where mountain-building forces were not as severe and the folds are, therefore, open and asymmetric and the faults are of the high-angle type. It is in the region of overlaps and extreme variations in type of deposition.

The first marked unconformity in the McCoy area is between Precambrian and Cambrian rocks. The break between Cambrian and Mississippian rocks, although representing a long time interval, is not conspicuous; it is a disconformity and no basal conglomerates were found in the area. The Mississippian-Pennsylvanian contact is also a disconformity but the surface is very irregular; channels and caves were carved in the Mississippian limestone before the Pennsylvanian sediments were spread over it. No break is recognizable from the bottom of the Pennsylvanian to the top of the Permian (?). The Triassic beds are separated from the underlying Permian (?) by a basal conglomerate but the beds of each system are parallel. There is a very slight angular unconformity between the Triassic and Jurassic formations. A lithologic change was the only indication of a break between Jurassic and Cretaceous beds observed within the area. Overlapping of the different formations has been discussed under "Sedimentary rocks" and is covered more fully by Lovering and Johnson (1933).

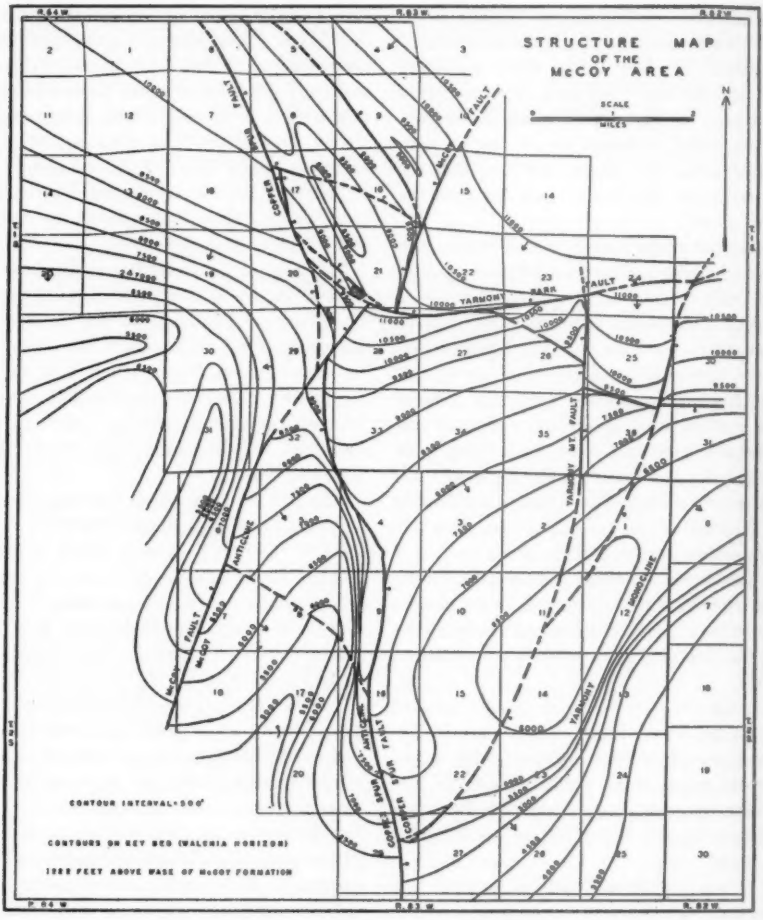

FIGURE 4.—STRUCTURE MAP OF THE McCOY AREA

Laramide structures

Folds.—The main folds trend north-south and are found in the southern half of the area (Fig. 4). They pitch to the south, in general. The Yarmony monocline passes through State Bridge and then northeast along the Colorado River, swinging around to the east after leaving the area. Dakota quartzites dip steeply to the southeast and in many places stand vertically along this fold (Fig. 5, section B–B′). West of Yarmony Mountain, the Copper Spur anticline passes about half a mile west of Bond in a northerly direction past the west side of Copper Spur beyond which it cannot be traced very far. This is an asymmetrical fold with the steep limb on the west side. The basal McCoy beds are slightly overturned at tunnel No. 43 on the Denver and Salt Lake railroad line.

The folds west of the Copper Spur anticline are smaller; in general the anticlines have a steep limb on the west side and a gentle one on the east side. Indications in this limited area are that the forces producing the folds came from the east.

In the northwest part of the area some small

folds (Fig. 6, section C–C′) trend slightly north of northwest. They are asymmetrical, with the steep limb of the anticlines on the northeast side. The largest one is found in Rock Creek valley between Crater and Volcano. It is slightly overturned and the exposed portion is well over 100 feet high.

The small folds are found in beds that strike approximately parallel to the axes of the folds. The beds, dipping southwest at 15-degree angle, have apparently been warped to give this regional dip by the elevation of the landmass to the northeast. The folds are believed to be drag folds formed by competent beds above riding up onto the flank of the highland during this warping.

Sediments of Niobrara age are involved in the folding. The next youngest rocks in the area are the Miocene lavas which are not folded but rest on the eroded surface of the largest anticline. The folds are, therefore, pre-Miocene and post-Niobrara. They are undoubtedly of Laramide age, but whether Montana, Paleocene, or Eocene cannot be told from evidence at hand.

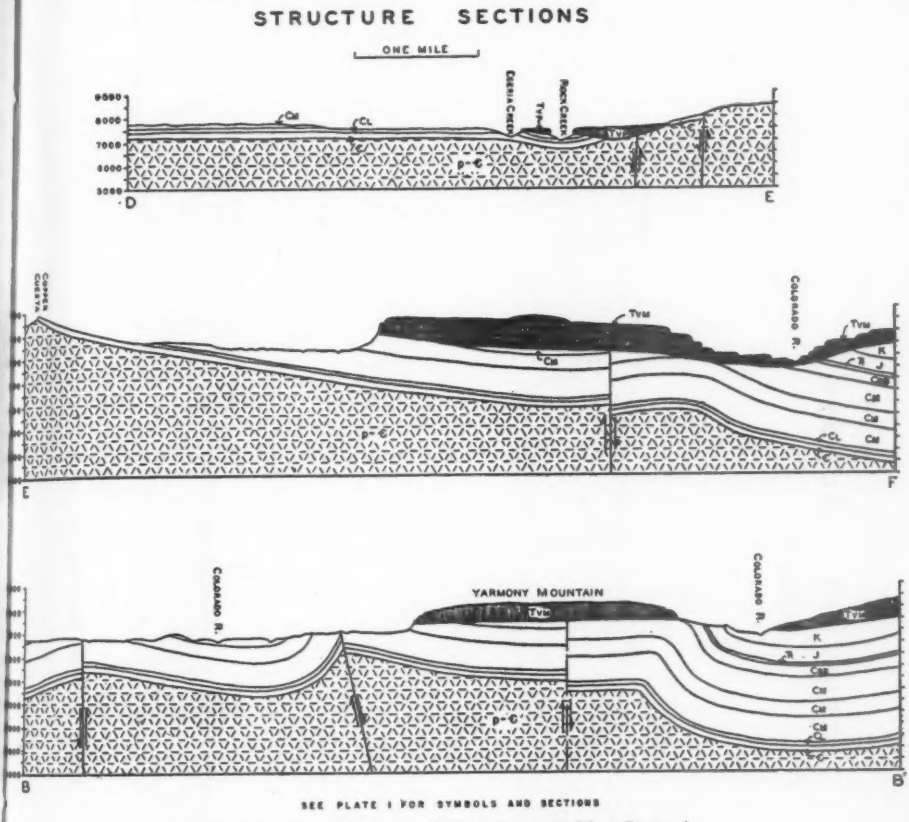

FIGURE 5.—CROSS SECTIONS D-E, E-F, AND B-B′ OF PLATE 1

Faults.—The fault pattern (Figs. 4, 5, 6) is that of a mosaic on a large scale and is interpreted to mean deformation due to vertical forces rather than horizontal. The main fault, the Copper Spur fault, enters the southern part of the area, passes west of Bond and Copper Spur, through Volcano, and leaves at the northwest corner. At Copper Spur the fault plane

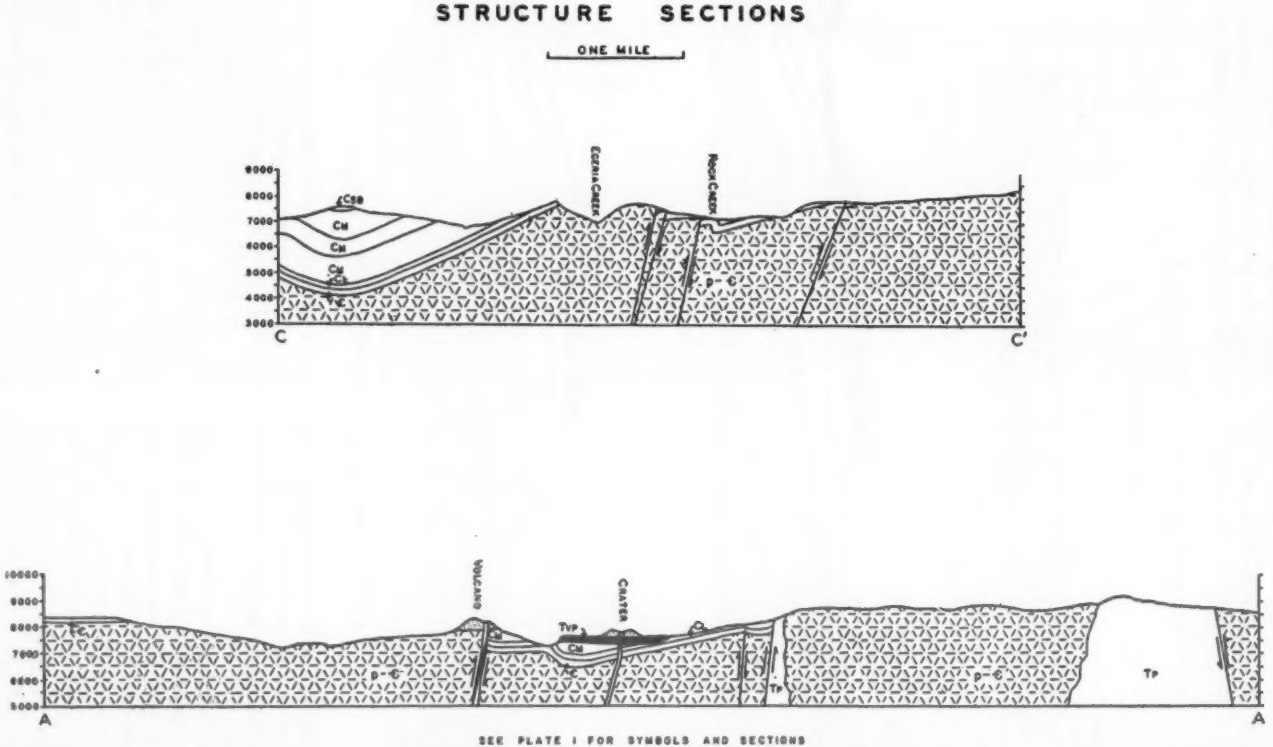

FIGURE 6.—CROSS SECTIONS C-C′, AND A-A′ OF PLATE 1

dips steeply northeast; northwest of Volcano it dips steeply southwest, where it is a reverse fault. In general, the faults are vertical or dip at steep angles and are normal faults usually showing considerable drag. South of a point about 1 mile northwest of Copper Spur, the fault is downthrown on the east side; north of this point it is downthrown on the west side until it intersects the McCoy fault, a north-northeast trending fault. The part of the fault which is downthrown on the east side cuts through the Copper Spur anticline (Fig. 5, section B–B′).

The fault north of Crater seems to have suffered reverse movement where it is exposed in Rock Creek valley (Fig. 7; Pl. 5, fig. 1). The drag shown indicates that the fault was at first a normal fault with drag on both sides. This fault probably formed when the region to the northeast was uplifted and the drag folds formed. Sag after the uplift changed the fault to a reverse fault, as it is today, with a vertical displacement of about 100 feet. In a stream valley about three-fourths of a mile southeast, this fault is again exposed. Here it has a displacement of over 200 feet. The Cambrian

FIGURE 7.—SUGGESTED MOVEMENTS ALONG FAULT PLANE IN PLATE 5, FIGURE 1

FIGURE 8.—CROSS SECTIONS C-C′, AND A-A′ OF PLATE 1

rocks, on the southwest or upthrown side, show no drag; the Pennsylvanian McCoy rocks, on the downthrown side, curve up almost vertically. From this exposure one would not suspect reversal of movement along the fault plane. The turned-up Pennsylvanian rocks are much less competent than the Cambrian quartzites, and, in sagging, there may have been a push to the southwest which turned up these rocks in the second movement regardless of how they were left after the first movement.

Reversed movement is also indicated along the Yarmony Park fault north of Copper Cuesta, but not as apparent as the previous case. In the southern part of section 22, T.1 S., R.83 W., Cambrian beds dip south at a 28-degree angle. A fourth of a mile south, across a fault with downthrown side on the north, Cambrian beds dip approximately south at a 40-degree angle with the dip decreasing rapidly to the south. This suggests a drag but for a fault downthrown on the south.

The positions of Crater and Volcano volcanic cones with respect to the faults show that the faults influenced their positions. The Tertiary intrusives were also influenced by the fault planes.

The way in which the faults cut through and displace the folds indicates that the faults are younger than the folds. This relation is well shown along the southern part of the Copper Spur fault (Fig. 5, section B-B′). The Miocene lavas are not cut by the faults, neither are the Tertiary intrusives. The faulting probably took place immediately after the major part of the folding had ceased and is, therefore, late Cretaceous or early Tertiary.

The main folds in this area could be drapings of the sedimentary rocks over faults in the Precambrian crystallines. The main faults appear closely related to the folds and are of minor importance when compared to them in size. The faults could be breaks in the later stage of vertical adjustments of Precambrian blocks. The relation between the Copper Spur anticline and fault (Fig. 5, section B-B′) would require a reversal of movement along the fault plane if this interpretation is correct. Reversal along other faults in this area has been pointed out.

Late Tertiary structures

The Miocene lavas of Yarmony Mountain and to the southeast have been tilted to a dip of about 9 degrees southwest. Those south of the area dip northeast. The region as a whole appears to have been folded into broad northwest-southeast trending folds after the lavas were deposited.

The Pliocene lavas east of Rock Creek have been elevated but not tilted. Their base is now over 400 feet above the Colorado River.

Geologic History and Physiography

An excellent summary of the geologic history of the Front Range and the regions immediately surrounding it is given by Lovering (1934, p. 21–23; 1935, p. 49–51). In his paper, "Geologic History of the Front Range, Colorado" (1929) this history is presented in detailed form. According to Lovering (1934, p. 21),

> "During pre-Cambrian time thick shaly sediments were deposited on an unknown basement, folded and metamorphosed into schists, intruded by large masses of granite, raised into mountains, and worn down to nearly featureless plains on both sides of an area of slightly higher land that existed along the site of the present Front Range."

In the McCoy area, this Precambrian complex consists mostly of the intruded granites. The surface over which the Cambrian sea spread was fairly even. The area mapped probably lies very near the northeast shore of that sea as Cambrian rocks thin out rapidly and disappear within a few miles when traced north and east of the area. This could, however, be due to erosion rather than lack of deposition. Slightly elevated Precambrian rocks to the northeast probably furnished the material for the Cambrian sediments.

During Ordovician, Silurian, and Devonian time there was no deposition in the area of this report, or what material was deposited was eroded before the Leadville limestone of Mississippian age was deposited. During this long interval the Cambrian sediments suffered no deformation, and the Leadville limestone was deposited parallel to the quartzites and shales. No pronounced elevation of the land

mass is indicated as the Mississippian sea was evidently clear. During late Mississippian and early Pennsylvanian time erosion was active, cutting channels and solution caves into the Leadville limestone and leaving a very irregular surface for the McCoy sediments of Pottsville age to be deposited upon. The Pennsylvanian sea spread further northeast than the earlier seas, overlapping the older formations. The coarse arkosic sandstones and conglomerates of Pennsylvanian age indicate a rising of the land mass immediately northeast of the Pennsylvanian sea in very early Pennsylvanian time. The interbedded, fossiliferous limestones and carbonaceous shales in the McCoy suggest that the shore line oscillated back and forth over the McCoy region; sometimes continental deposits were formed and other times marine deposits. Sudden changes in climatic conditions must also have taken place. The disappearance of the black shales, the diminishing of the limestones and micaceous shales and the increasing redness of the beds indicate that the climate was becoming more arid during late Pennsylvanian and Permian (?) time. Pennsylvanian time passed into Permian (?) without crustal movement or changing climate, and no break in stratification is noticeable. The conglomerates and coarse sandstones become finer and brick-red shales and siltstones take their place. This indicates that the Front Range highland was becoming worn down and the stream gradients were decreasing.

A break in sedimentation is indicated between Permian (?) and Triassic time by a conglomerate with fragments of fossil wood at the base of the Triassic beds. Shales and mudstones follow. The Triassic sea was not as extensive as the Permian (?) sea, however, and was of short duration. Sedimentation and erosion were light during the Triassic. An angular unconformity of small discordance is suggested between the Triassic and the overlying Jurassic beds, indicating a slight tilting up in the northeast.

The Front Range Highland had probably been reduced to a series of low hills by the Jurassic, and the cross-bedded Entrada sandstone and the ripple-marked sandstones and shales of the Morrison with their vertebrate fossil remains were deposited on a broad piedmont plain. By mid-Cretaceous time the highland was almost completely submerged and the Dakota sandstones were deposited over practically the whole of Colorado. The highland remained submerged until after Pierre time when it started once more to rise above the sea and form the present Front Range. The rocks of the McCoy region were folded and faulted during the Laramide revolution; in early Tertiary time they were intruded by porphyries. In the Eocene and Oligocene epochs, erosion, predominant in this region, carried the material to the southwest.

Volcanic activity in the Miocene produced basaltic and andesitic lavas with interbedded tuffs, breccias, and ash. The erosion surface upon which this lava was deposited may possibly be correlated with the Bergen Park erosion surface of Van Tuyl and Lovering (1935, p. 1317–1321). It is here called the Black Mountain surface. Marked uplift in the Front Range after the lavas were deposited warped them up northeast of State Bridge; in the region of State Bridge and Piney Creek they sagged. Erosion continued to dissect the region. In Pliocene time, lava flows poured down a stream valley north of McCoy. Two volcanic cones of scoria and ash were later built upon these flows.

Pediments, formed probably in late Pliocene and early Pleistocene time, are cut across deformed rocks and are covered with a thin coating of loose material. Good examples may be seen in the northeast corner of section 8, T.2 S., R.83 W. and in the northeast corner of section 17, T.2 S., R.83 W. It is generally believed that arid conditions prevailed here during late Pliocene and early Pleistocene. Sheet erosion might therefore have been pronounced and instrumental in forming the pediments. The upper or rock terrace at Bond, and others of about the same level, probably represents the river level when these pediments were formed.

A broad regional uplift in early Pleistocene time rejuvenated the streams, and the Colorado River cut down its channel to within probably about 25 or 30 feet of its present level. This is indicated by the fact that in general the gravel in the terraces does not extend below this level in the area. Gravel fills below the present

river bed have been reported outside of the area.

The mountains to the northeast were glaciated. When these glaciers began to retreat, the streams were probably overloaded with debris and the river began to aggrade. Coarse gravels were deposited first; as the volume of water coming from the glaciers decreased and the streams slowed up, the finer material was deposited. The height of these gravel and silt terraces is approximately 150 to 200 feet above the present level of the river. The rock terrace is some 50 feet higher. With the passing of the glaciers, the streams cleared up and began to cut new channels into these deposits. The river has now cut 25–30 feet below the level of the original channel.

In the southwest corner of section 5, T.2 S., R.83 W., the Colorado River has altered its course somewhat. A meander, formed before the gravels were deposited, was cut off when the river cut through the gravels and reached bedrock again.

Wind gaps due to stream capture in the smaller streams were found immediately outside the area. One of these forms the gap in the ridge west of Piney Creek about 2½ miles south-southeast of State Bridge. Another is found in the Dakota hogback about 1½ miles south of Red Gorge.

Three erosional surfaces were recognized besides the river terraces. These surfaces will be further discussed in a forthcoming paper by the author on the physiography of the Colorado River valley from Gore Range to Dotsero.

The Black Mountain surface is the highest and oldest. It is named from Black Mountain, north of the McCoy area, which probably stood as a monadnock above this peneplane. This is the surface upon which the Miocene lavas of Piney Ridge and Yarmony Mountain were poured out. It can be traced across the Gore Canyon from Black Mountain to Sheephorn Mountain.

The Black Mountain surface was tilted and folded into broad warps or folds after the Miocene lavas had been deposited. The Sheephorn surface, well developed between Sheephorn Mountain and Sheephorn Creek, was developed on this uneven surface. It represents a considerable period of base leveling because the surface is eroded rather uniformly over Precambrian and all younger sedimentary rocks. It is recognized north of Yarmony Mountain and along its east slope, on the north slope of Piney Ridge and as shoulders on the slopes above Hartman divide and on the east nose of Red Canyon.

Elevation of the Sheephorn surface rejuvenated the area and the Blacktail surface was developed about 500 feet below the Sheephorn. It is well developed where the Colorado River cuts through it to form the Blacktail Canyon north of Radium. It is also recognized at Red Canyon, Yarmony Station, Piney Ridge, and a number of other localities. The Colorado River has cut some 500 feet below this surface. It must therefore be higher and older than the surface upon which the late lavas flowed down Rock Creek. It must be pre-Pleistocene.

Economic Geology

Copper and lead

About 2 miles up Yarmony Creek from Copper Spur is Cole's mine, from which copper and lead ore have been taken. The following history of the mine was obtained from a preliminary report written for E. J. Longyear Company and lent to the writer by that Company (Schwartz, 1930), supplemented by a written communication in 1936 from Mr. K. R. Wyman, then postmaster of Copper Spur.

The deposit was originally located by Frank Graw of McCoy who, with others, sank an incline on the ore and then sold out for $15,000. From 1914 to 1917 the mine was operated by the Copper King Mines Products Company which erected the leaching and electrolytic plant at Copper Spur and extended the incline to about its present length. One carload of copper plates was shipped from Copper Spur, but operations ceased in 1917 because of lack of funds. In 1921, the company reorganized under the name Electro Copper Company but no ore was mined. In 1929 the Cole Mining and Milling Company obtained a lease on the property with option to purchase. They installed two old-style Hartz jigs and two Wilfley tables and expected to ship high-grade concentrates. Both copper and lead concentrates were obtained, but they ran out of funds before pro-

ducing enough for a shipment. Mr. Wyman estimates that at least $150,000 has been invested in the property. The mill has now been completely dismantled.

A dirt road connects the mine with the main road. The mine is opened by an incline and drifts; the total length of openings is over 1000 feet. The dip varies from 12 to 15 degrees. The workings extend down dip in general and follow the conspicuous green and blue staining due to malachite and azurite which is the main evidence of mineralization. This staining seems to follow a rather definite bed of sandstone below a shale unit. The deposit occurs in sandstone of the Sawatch formation only a few feet below its contact with the Leadville limestone of Mississippian age. According to Schwartz (1930), the mineralization varies from 2 to 12 feet in thickness, and at a few places unoxidized nodules occur which are predominantly a galena sandstone, the galena occurring as the cement. With the galena are lesser amounts of chalcopyrite and pyrite. There is a limited amount of carbonate cement which, it is believed, has been largely replaced by the sulphides. Schwartz reports gold, silver, and zinc from this mine as well as copper and lead. His report (1930) states:

"It seems evident . . . that the metallic minerals were introduced into the sandstone, replacing the carbonate cement and possibly in some cases attacking the sand grains. The original minerals were probably sulphides as indicated by unoxidized nodules found in the mine. Oxidation has been active in the surface and near surface areas of the porous sandstone. The amount of the metallic minerals present seems to vary with the original porosity of the sand. Where the porosity was high the amount of cement was greater, and therefore, more copper and lead minerals could get in."

An assay map included in Schwartz's report gives the estimated ore in sight as 9000 tons at 2.71 per cent copper and the balance of the mine at 1.16 per cent copper. The lead content had not been recognized when this work was done.

When Schwartz's report was written the sandstone in which the mineralization occurs was believed to be Upper Carboniferous. Detailed mapping has shown that it is probably Cambrian. It is the writer's opinion that mineralization was due to the deeper portion of the Tertiary porphyry intrusives which are exposed 2 miles north and 2½ miles northeast of the mine but may also underlie this area. The shale bed and limestone above the mineralized zone apparently acted as a barrier to rising emanations from the intrusives. In the Leadville area the Leadville formation is a dolomite, and the mineralization took place in it instead of below it. It is suggested that, in the contraction due to the changing into dolomite, fine openings were formed in the Leadville which allowed the emanations to enter, while at Copper Spur the pure limestone acted as a barrier. The deposits were probably formed at moderate depths by ascending hot solutions.

Prospect pits and adits are scattered throughout the area wherever the Cambrian-Mississippian contact is exposed. Experience has led the prospectors to expect to find deposits at this horizon in the McCoy area.

Lime

The Leadville limestone is very pure in places and has been used for making quicklime. A small abandoned kiln is located about a fourth of a mile northeast of Crater. The post office at Volcano Siding is called Hydrate, and the limestone a fourth of a mile south of it has been worked to some extent.

Ballast

Perhaps the most important economic product produced in the McCoy area is the ballast quarried from Volcano. Thousands of carloads of scoria have been removed from the east side of this cone and used for ballast on railroad lines. A small settlement of old railway coaches and box cars has grown at the cone to accommodate the laborers and their families.

Mineral springs

A fourth of a mile southeast of McCoy is a mineral spring in the form of a pool about 20 or 25 feet in diameter which has built a rim of calcareous tufa around it. This rim continues to build up as the water flows over the lip. The water is warm but not hot, possibly about 70 degrees F. A number of minerals have been

detected in it according to local reports. Cattle drink the water.

In the southwest corner of section 16, T.2 S., R.83 W. is another spring which trickles out of Cambrian rocks near the bottom of the hillside. The water tastes very bad and near the spring the ground and vegetation over which the water flows is covered with a red slimy deposit. Farther away a white crust covers the ground.

The occurrence of both springs close to large faults near the Cambrian-Mississippian contact suggests a relationship.

Summary

The Precambrian rocks of the McCoy area are largely granite with local areas of schist and gneiss. Sedimentary rocks of Ordovician, Silurian, and Devonian age are missing. Because of fossil evidence and lithologic character, limestones assigned by others to the Ordovician system are herein recognized as the Leadville limestone of the Mississippian system. The McCoy formation is extended to include all the rocks from the top of the Leadville limestone to the base of the State Bridge siltstone. This includes over 3500 feet of coarse arkosic continental deposits with interbedded shales and limestones. The Walchia bed is introduced as a member of this formation, which is of Des Moines age. The term State Bridge siltstone is applied to the brick-red siltstones overlying the McCoy formation. Its age is middle or late Pennsylvanian or Permian. Sedimentary rocks of Triassic, Jurassic, and Cretaceous age are also represented.

The area was folded into open asymmetric folds in late Cretaceous time and then faulted. Some of the faults have a displacement of several hundred feet. The faults are mainly of the high-angle, normal type. Mineralization in the area is interpreted as being due to the Tertiary intrusives of monzonite porphyry which are found in the northeastern part of the area. On the Geologic Map of Colorado (U. S. Geol. Survey, 1935) these intrusives are shown as Precambrian granite. Basaltic and andesitic lavas with interbedded tuffs, breccia and ash were formed during Miocene time. A much smaller basaltic flow formed during Pliocene time with the building of two cinder cones at a somewhat later date.

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NEW HAMPSHIRE STATE PLANNING
AND DEVELOPMENT COMMISSION AND HIGHWAY DEPARTMENT

NEW HAMPSHIRE
MONADNOCK QUADRANGLE

LEGEND

METAMORPHIC AND IGNEOUS ROCKS

LATE DEVONIAN? — *New Hampshire Magma Series*

Concord granite
(Light-gray fine-grained to medium-grained massive to foliated granite, quartz monzonite, or granodiorite composed of potash feldspar, oligoclase, quartz, biotite, and some muscovite; locally has phenocrysts of potash feldspar.)

Spaulding quartz diorite
(Dark-gray medium-grained massive to well-foliated granodiorite and quartz diorite composed of oligoclase-andesine, potash feldspar, quartz, and biotite; aggregates of the light and dark minerals give the rock a spotted appearance.)

Kinsman quartz monzonite
(Porphyritic coarse-grained dark-gray quartz monzonite, granodiorite, and quartz diorite composed of potash feldspar, oligoclase, quartz, biotite, and muscovite; phenocrysts of orthoclase or microcline rarely oligoclase-andesine; phenocrysts commonly form augen.)

Amphibolite
(Area where amphibolite sills and dikes are abundant.)

MIDDLE DEVONIAN? — *Oliverian Magma Series*

Quartz diorite and granodiorite
(Medium-grained gray to pink subporphyritic weakly foliated granodiorite or quartz diorite, composed of oligoclase-andesine, potash feldspar, quartz, biotite and muscovite.)

LOWER DEVONIAN

Littleton formation
(Bulk of formation (Dl) is composed of mica schist, quartz-mica schist, sillimanite schist, and sillimanite-garnet schist. Concretions of actinolite granulite, garnetiferous quartzite, and quartzite occur in some of the schist. Rusty quartzite member (Dlr) composed of pyritiferous rusty-weathering quartzite, lime-silicate granulite, dark-colored fine-grained granular quartz-biotite schist, and mica schist. Upper part of formation (Dlg) is gneiss and rusty-weathering schist; concretions occur in many places.)

SILURIAN — *Lower or Middle*

Clough quartzite
(Compact white to bluish-gray quartzite, quartz conglomerate, and some thin beds of quartz-muscovite schist and quartz-garnet-muscovite schist.)

VICIAN? — *...per?*

Partridge formation
(Mica schist, mica-garnet schist, sillimanite-garnet schist, and micaceous quartzite. The Partridge formation is generally present between the Clough and Ammonoosuc formations but is too thin to show on the map.)

ORDOVICIAN? Upper?

Op

Partridge formation

(Mica schist, mica-garnet schist, sillimanite-garnet schist, and micaceous quartzite. The Partridge formation is generally present between the Clough and Ammonoosuc formations but is too thin to show on the map.)

Oam

Ammonoosuc volcanics

(Light-colored fine-grained quartz-feldspar-biotite gneiss; dark-colored amphibolite and amphibolite gneiss.)

METAMORPHIC ZONES

All metamorphism is high-grade

CONTACTS

Accurate

Approximate and diagrammatic due to poor exposures.

Indefinite as sharp contact is lacking.

SPECIAL SYMBOLS

(Strike and dip symbols represent only a small percentage of the field observations.)

Strike and dip of bedding.

Strike of vertical beds.

Horisontal bedding

Strike and dip of foliation and schistosity

Strike of vertical foliation and schistosity

Silicified zone

Pegmatite

Granite quarries, mostly abandoned, only the largest are shown

G

Abandoned gold mine

Gr

Abandoned graphite prospect

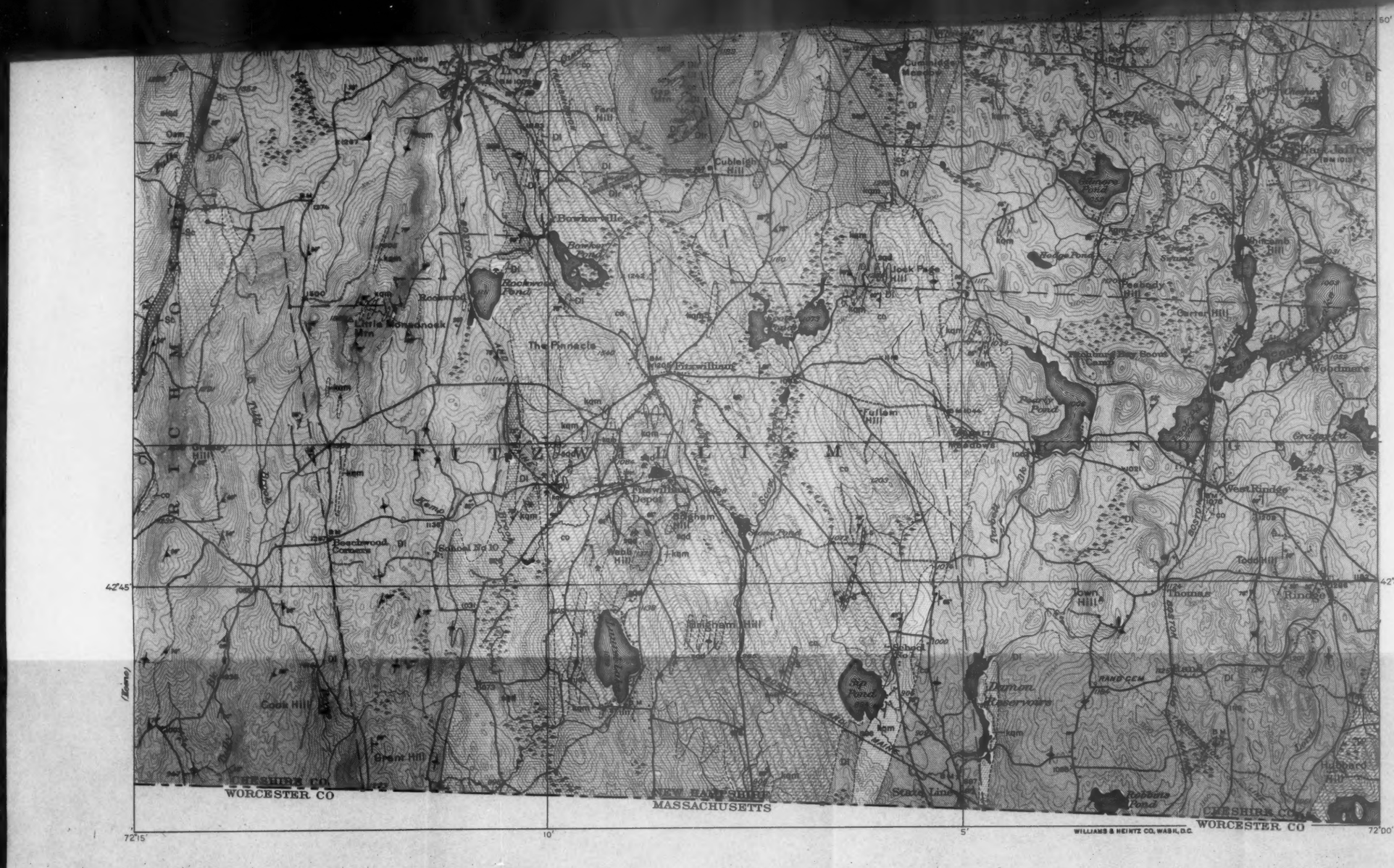
SECTION A-A′

SECTION B-B′

SECTION C-C′

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE MONADNOCK QUADRANGLE, NEW HAMPSHIRE

Geology by Katharine Fowler-Billings assisted by L. Nolan, D. Mack, Mr. R. Chaisson. Geology surveyed in 1941, 1942,


Cunningham Meadow
Bowkerville
Bowker Pond
Rockwood Pond
Little Monadnock Mtn.
The Pinnacle
Fitzwilliam
Cubleigh Hill
Fern Hill
Cook Page Hill
Tullam Hill
Peasley Pond
Peasley Hill
Carter Hill
Whitcomb Hill
Woodmere
Gilsum Rd.
West Ridge
Todd Hill
Thomas Ridge
Town Hill
Depot
Brigham Hill
Webb Hill
Beechwood Corners
School No 10
Coats Hill
Grant Hill
Dumpling Ridge
Rand Gem
Rabbit Pond
Hubbard Hill
Scale Line
Boston
Main
Ridge
Road
Brook
Creek
Pond
Hill
Mtn.
Corners
Depot
Ridge
Reservoirs
Meadows
Woods

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GEOLOGY OF THE MONADNOCK REGION OF NEW HAMPSHIRE

By Katharine Fowler-Billings

CONTENTS

	Page
Abstract	1250
Introduction and acknowledgements	1250
Topography and drainage	1252
Glacial features	1252
General lithologic features	1253
Ammonoosuc volcanics	1255
Partridge formation	1255
Clough quartzite	1257
Littleton formation	1257
General statement	1257
Lower member of Littleton formation	1258
Rusty quartzite member	1258
Middle member of Littleton formation	1261
General statement	1261
Mica schist	1261
Sillimanite schist and sillimanite-garnet schist	1261
Impure quartzite	1262
Rocks with retrograde minerals	1262
Upper member of Littleton formation	1263
Nodular concretions in Littleton formation	1264
Oliverian magma series	1264
New Hampshire magma series	1265
Kinsman quartz monzonite	1265
General statement	1265
Megascopic description	1265
Microscopic description	1265
Correlation	1266
Spaulding quartz diorite	1266
General statement	1266
Megascopic description	1266
Microscopic description	1267
Correlation	1268
Concord granite	1268
General statement	1268
Megascopic description	1269
Microscopic description	1269
Correlation	1269
Dikes and sills	1270
Amphibolite dikes	1270
Pegmatites and quartz veins	1270
Biotite schist dike	1271
Structure	1271
General statement	1371
East limb of Swanzey dome	1271
Folded rocks of Littleton formation	1272
Mt. Monadnock syncline	1272
Evidence of rusty quartzite member	1274
Rest of Littleton formation	1275
Thickness of Littleton formation	1275
Igneous bodies	1275
Oliverian magma series	1275
New Hampshire magma series	1275
Faulting—silicified zones	1277
Age of structural features	1278
Summary of geologic history	1278
References cited	1279

ILLUSTRATIONS

Figure	Page
1. Index map showing location of Monadnock quadrangle and recently mapped adjacent quadrangles	1251
2. Columnar section for Monadnock area	1254
3. Section of rusty quartzite along brook 1¼ miles southwest of Dublin Pond	1260
4. Thrust contact of granodiorite of Oliverian magma series and Ammonoosuc volcanics	1272
5. Block diagram of folds	1275

Plate	Facing page
1. Geologic map and structure sections of the Monadnock quadrangle	1249
2. Mt. Monadnock rising above the New England Upland	1252
3. Rocks and structures in Monadnock region	1253
4. Pseudo-sillimanite crystals	1262
5. Photomicrographs	1263
6. Structure map	1272

TABLES

Table	Page
1. Measured sections of the Ammonoosuc volcanics and Clough quartzite	1256
2. Approximate modes—Ammonoosuc volcanics, Partridge formation, and Clough quartzite	1257
3. Approximate modes—Littleton formation	1259
4. Approximate modes—granodiorite of oliverian magma series	1265
5. Approximate modes—Kinsman quartz monzonite	1267
6. Approximate modes—Spaulding quartz diorite	1268
7. Approximate modes—Concord granite	1269

Abstract

Mt. Monadnock, the dominant physiographic feature of the Monadnock region, rises 2000 feet above the New England Upland in southwestern New Hampshire and is the type example of a monadnock. The region is underlain by metamorphic and plutonic rocks that range in age from Ordovician (?) to late Devonian (?).

The metamorphic rocks of the area belong to four formations. The Ordovician (?) Ammonoosuc volcanics, consisting of light-colored, fine-grained biotite gneiss and dark amphibolite, attain a maximum thickness of 600 feet in this region. The Ordovician (?) Partridge formation, chiefly mica schist, is not more than 50 feet thick in this area. The Silurian Clough quartzite here attains a maximum thickness of 100 feet. Much of the area is underlain by the lower Devonian Littleton formation which is 17,000 feet thick in the Monadnock area. The Littleton formation can be divided into several members. A lower member, 5000 feet thick, is composed of mica schist, impure quartzite, and gneiss. Above this is the rusty quartzite member which does not exceed 600 feet in thickness and is composed of pyritiferous rusty-weathering quartzite, actinolite granulite, quartz-biotite schist, and mica schist. It has proved to be an indispensable tool in unraveling the stratigraphy and structure. Above the rusty quartzite member is the middle member, 5400 feet thick, characterized by excellent bedding and composed of mica schist, sillimanite schist, sillimanite-garnet schist, and impure quartzite. Mt. Monadnock is held up by this resistant middle member of the Littleton formation. The upper member of the Littleton formation, about 6000 feet thick, is composed primarily of a massive, rusty-weathering gneiss.

Metamorphism is high grade throughout the area, although some of the rocks contain retrograde minerals typical of lower metamorphic zones.

Half of the Monadnock region is underlain by plutonic rocks. The oldest intrusive rocks are granodiorite and quartz diorite which belong to the Oliverian magma series of Devonian (?) age. The intrusives of the New Hampshire magma series—the Kinsman quartz monzonite, the Spaulding quartz diorite, and the Concord granite—are late Devonian (?). Pegmatites are associated with each of the intrusives.

The east limb of the Swanzey dome occupies the west half of the area, and the Monadnock syncline is the most prominent structural feature in the central part of the area. The rusty quartzite member zigzags across the central part of the region in a series of anticlines and synclines and is the key horizon for determining the structure and thickness of the middle part of the area. The northern part of the region is made up of gneiss; the southern part of the area is so thickly covered by glacial drift that the structure is difficult to determine.

The Oliverian magma series, exposed along the western margin of the Monadnock region, appears to be part of a large concordant sheet injected near the top of the Ammonoosuc volcanics. The Kinsman quartz monzonite, apparently concordant in other parts of New Hampshire, is cross-cutting and of batholithic proportions in the northeastern part of the Monadnock region, but elsewhere in the area occurs as small concordant and discordant bodies. The Spaulding quartz diorite forms sill-like masses as well as discordant bodies; it makes up the core of an anticline in the Stone Pond region. The Concord granite is cross-cutting.

The region was folded in middle or late Devonian time. Intrusion of the Oliverian magma series preceded the folding. The New Hampshire magma series was intruded after the sediments were folded, but before deformation had ceased. Late faulting following the deformation and igneous intrusion is indicated by the presence of silicified zones.

Introduction and Acknowledgments

The Monadnock region as defined for this paper includes the Monadnock quadrangle, New Hampshire, and the adjoining part of the Winchendon quadrangle as far south as the Massachusetts State line (Fig. 1). Special attention is paid to the bedrock geology with brief reference to physiographic and glacial features. C. T. Jackson (1844) and C. H. Hitchcock (1874; 1877; 1878) mention Mt. Monadnock briefly. Emerson (1917) describes some of the formations found in the extreme southern part of the Monadnock region. Wheelock (1873) studied the glacial striae on Mt. Monadnock, and Gulliver (1899) described a few of the physiographic features of the region. Using Mt. Monadnock as the type example, W. M. Davis (1896, p. 192) first applied the term "monadnock" to a residual hill or mountain rising above a peneplain. Except for a paper by Perry (1904), no detailed description of the geology of the region has been published.

The field work in the Monadnock quadrangle occupied the entire summers of 1941, 1942, and 1943; in addition to general geological work, special studies were made of pegmatites and sillimanite deposits in the area. The State

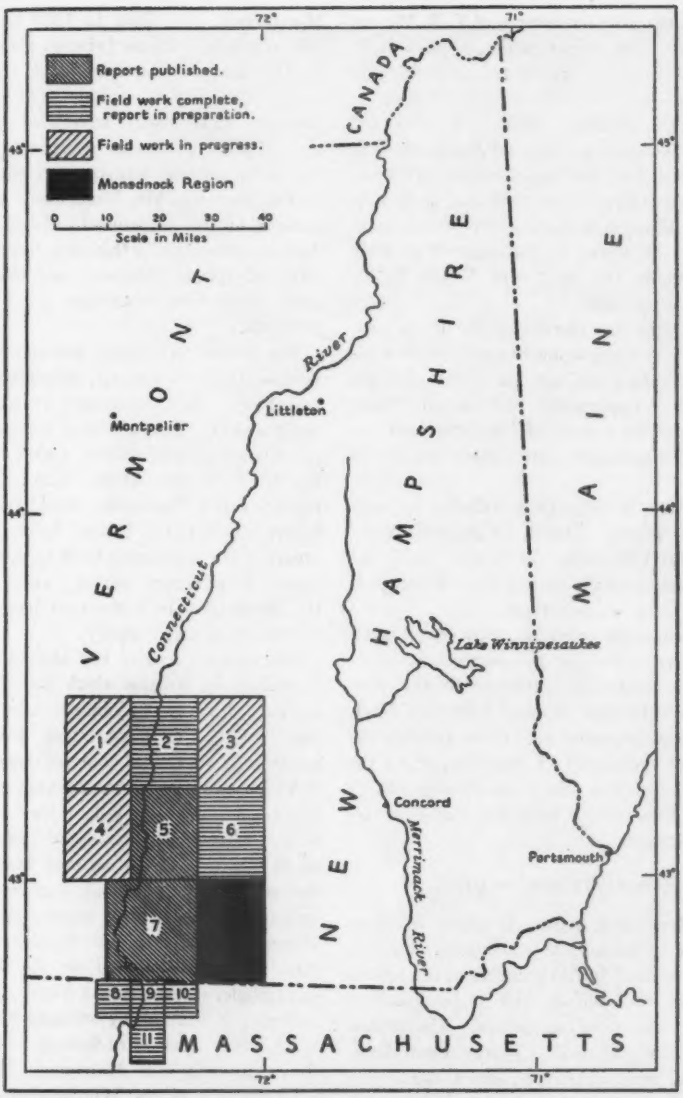


FIGURE 1.—INDEX MAP SHOWING LOCATION OF MONADNOCK QUADRANGLE AND RECENTLY MAPPED ADJACENT QUADRANGLES

1, Ludlow (J. B. Thompson); 2, Claremont (R. W. Chapman and C. A. Chapman; see also R. W. Chapman and C. A. Chapman, 1940 and C. A. Chapman, 1942); 3, Sunapee (C. A. Chapman); 4, Saxtons River (J. L. Rosenfeld); 5, Bellows Falls (F. C. Kruger, 1946); 6, Lovewell Mountain (M. T. Heald); 7, Keene and eastern third of Brattleboro (G. E. Moore, 1949); 8, Bernardston (R. Balk); 9, Northfield (R. Balk); 10, Mt. Grace (J. B. Hadley); 11, Millers Falls (R. Balk).

Planning and Development Commission of New Hampshire sponsored and contributed funds for the expenses of the sillimanite investigation as part of their war-time strategic mineral surveys. The writer is indebted to Marland P. Billings who made several visits to the area and

who, with the added assistance of T. R. Meyers and N. K. Flint, helped with the plane-table surveys of the Gap Mountain and Cobb Hill sillimanite deposits described in an earlier report (Fowler-Billings, 1944b). Thanks are also due Leo Nolan and Donald Mork who were field assistants in this earlier work. To complete the mapping of the area and to include the New Hampshire part of the Winchendon quadrangle, 6 weeks of the summer of 1946 were spent in the field with Ursula Bailey Chaisson as assistant.

The geology was plotted on the U. S. Geological Survey topographic maps of the Monadnock quadrangle (edition of 1936) and the Winchendon quadrangle (edition of 1935). These maps, on a scale of 1/62,500, were enlarged photostatically three times for use in the field.

Laboratory investigations included a study of 82 thin sections. The Shaler Memorial Fund of Harvard University contributed funds for the completion of the project in 1946 and paid for most of the thin sections.

The photographs, unless otherwise acknowledged, were taken by the writer. Some of these were originally published in the New England Naturalist (Fowler-Billings, 1941). The photomicrographs were taken through the courtesy of Professor C. S. Hurlbut, Jr., of the Department of Mineralogy and Petrography of Harvard University. Drafting was done by Edward Schmitz.

Topography and Drainage

The Monadnock region is about 70 miles northwest of Boston, Massachusetts (Fig. 1). It can be reached by two branches of the Boston and Maine Railroad as well as by excellent highways with good bus service. The largest villages in the area are East Jaffrey, Fitzwilliam, Troy, Marlboro, Harrisville, and Dublin.

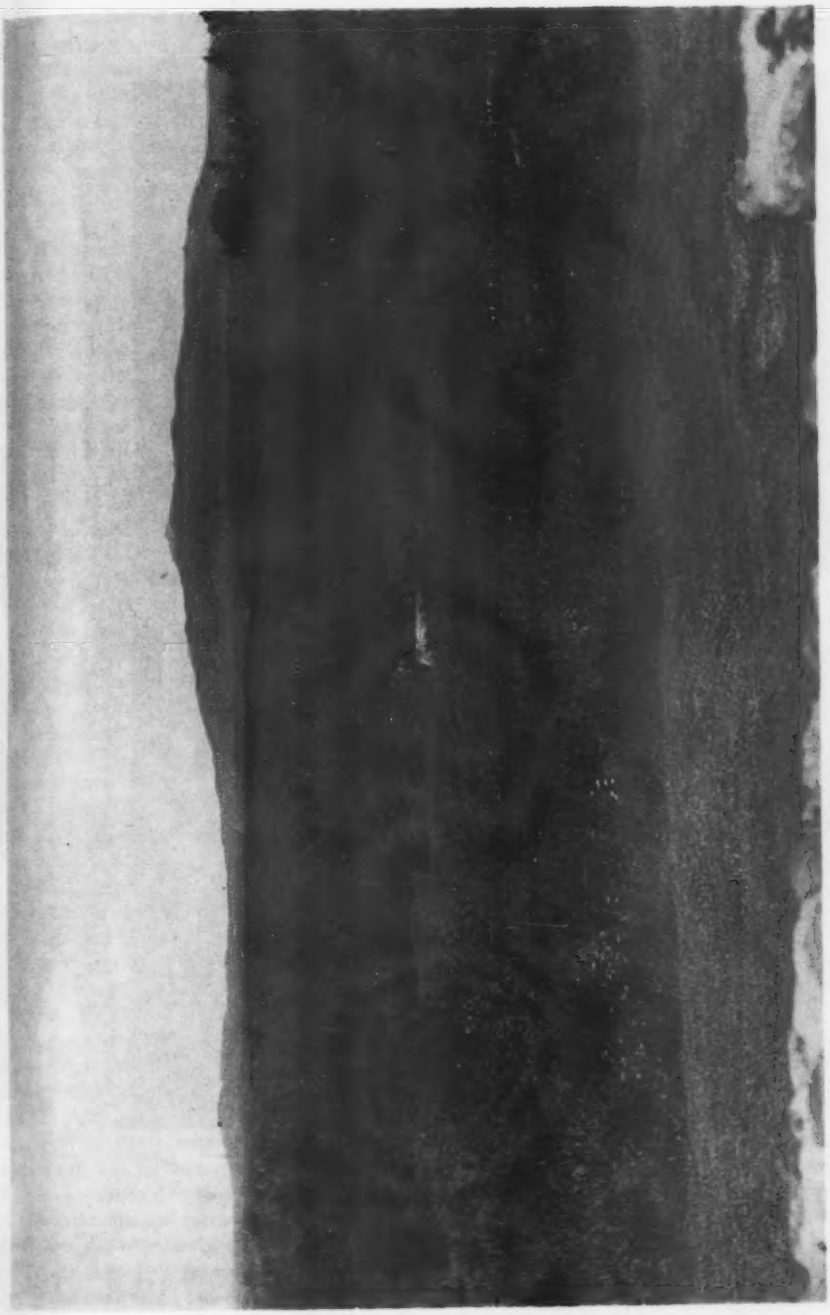
Mt. Monadnock, the dominant physiographic feature of the region, 3165 feet in altitude, rises 2000 feet above the surrounding lowlands (Pl. 2). W. M. Davis (1909) and D. W. Johnson (1931, p. 1–22) discuss fully the age of the dissected erosion surface called the "New England Upland" above which Mt. Monadnock rises. The upland is a dissected peneplain now elevated to 1000 to 1200 feet south of Mt. Monadnock, but rising to 1400 feet north of the mountain. Davis believed the New England Upland to be an extension of the Schooley peneplain which he considered to be Cretaceous; Johnson subsequently advocated a Tertiary age. Uplifting of this peneplain caused rejuvenation of the drainage and entrenchment of the streams. Mt. Monadnock is the most striking of the monadnocks rising above the Tertiary peneplain in this area, but Skatutakee (2002 feet), Gap (1862 feet), and Little Monadnock (1883 feet) mountains are also similar residuals.

The Monadnock region has numerous lakes, most of which are natural, although many have had their levels raised slightly by small dams at their outlets. This is true of Nubanusit Lake, Spoonwood Pond, Silver Lake, Woodward Pond, Harrisville Pond, Skatutakee Lake, Dublin Pond, Thorndike Pond, Pearly Pond, Coontoocook Lake, Laurel Lake, and many others. The dams were built to develop water power, to eliminate swampy areas by raising the shore lines to a constant level, for flood control or for water supply.

The western half of the Monadnock region is drained by streams which flow in a general westerly direction to join the Ashuelot River, and finally the Connecticut River. Their headwaters are typical youthful streams flowing in V-shaped valleys entrenched several hundred feet below the general upland level. The gorge of the Ashuelot River north of Troy, the gorge of Minnewawa Brook east of Marlboro, and the gorge of Otter Brook southwest of East Sullivan are the three main outlets of the streams and lakes of the western half of the area. The eastern half of the Monadnock quadrangle, except Dublin Pond which drains westerly, is drained by streams which finally reach the northward-flowing Coontoocook River to join the Merrimack River. The northern part of the Winchendon quadrangle drains southward by a series of small brooks.

Glacial Features

The Monadnock region shows many features resulting from Wisconsin glaciation. The ice pushed across the area in a southeasterly direction, and the topography was modified; streams were clogged with till, sand, and gravel;



MT. MONADNOCK RISING ABOVE THE NEW ENGLAND UPLAND

View was taken from Sharon, 10 miles east-southeast of Mt. Monadnock. Upland level is 1100–1200 feet in elevation. Photo by Mrs. Bernice B. Perry.

FIGURE 1. SUMMIT OF MT. MONADNOCK
As viewed from the south.

FIGURE 2. QUARRY IN CONCORD GRANITE
1¼ miles south of the village of Marlboro, east of highway.
Photo by E. Wm. Heinrich.

FIGURE 3. VIEW FACING NORTH OF EAST-DIPPING BEDS OF LITTLETON FORMATION ON WEST LIMB OF MT. MONADNOCK SYNCLINE
0.1 mile west of summit of Mt. Monadnock; light beds are quartzite; gray beds are mica schist.

FIGURE 4. FOLD IN LITTLETON FORMATION
0.1 mile west of summit of Mt. Monadnock.

bedrock was scoured and exposed or deeply covered by drift.

Mt. Monadnock, the highest point in the path of the ice, was cleaned of soil. The bedrock on the summit and northern slopes was smoothed and polished into *roches moutonnées*, while the southern slopes suffered plucking, leaving jagged cliffs controlled by joints (Pl. 3, fig. 1). The average direction of glacial movement on the mountain was S.20°E. Deflection of the ice by the mountain caused this direction to vary in many instances. In the vicinity of the Halfway House, the striae trend S.65°E. On the road a few hundred yards below the Halfway House, two sets of striae were formed, the underlying set trending S.65°E. and the upper set trending S.40°E. Near the summit of Monte Rosa, the 2480-foot knob 0.5 mile N.10°W. of the Halfway House, two sets of striae were also observed, one set trending almost due east and the other set trending S.40°E.

The present bareness of the summit and ridges of Mt. Monadnock is attributed to forest fires which have burned off much of the thin postglacial humus. A few stone nets formed at the end of glacial times on the higher slopes of the mountain. These are best exposed in the flats around 3000 feet along the Pumpelly Ridge Trail, where grassy centers a few feet across are surrounded by small boulders pushed out by frost action.

North of Mt. Monadnock the average trend of the striae is S.20°E. but ranges from S.10°E. to S.40°E. South of the mountain the striae have similar trends, but lunoid furrows along the shores of Sip Pond and striae near Rockwood Pond indicate that the ice moved south in some places.

The whole area east, southeast, and south of Mt. Monadnock, as far south as the Massachusetts State line, is an area of drumlins and outwash with only occasional exposures of rock. Possibly, when passing over Mt. Monadnock, the ice sheet became saturated with debris which was then dumped on the lee side of the mountain as drumlins. The stagnating ice, during the disappearance of the continental glacier, was full of material that was deposited as sand and gravel by the meltwater of the ice sheet. Consequently, swampy areas, lakes, and sluggish streams are common in this southeastern part of the area.

The region to the north, northwest, west, and southwest of Mt. Monadnock as far south as the Massachusetts State line is generally covered by till with scoured bedrock hills and few drumlins; sand and gravel clog the valleys. Occasional eskers and recessional moraines are present. The 3¼-mile section of Otter Brook north of South Keene is in a broad preglacial valley which has been filled by sand and gravel, so that Otter Brook meanders in a wide valley, rarely reaching bedrock. Large potholes were scoured by glacio-fluvial streams in the bedrock almost 100 feet above the present channel of Otter Brook on the north side of the east-west road just north of Roxbury School. The stretch of river 1 mile east of South Keene is also in a sand-filled valley; the river has not yet cut through to bedrock.

General Lithologic Features

Over 50 per cent of the Monadnock area is composed of highly metamorphosed sediments of the Littleton formation of Devonian age They include gneiss, mica schist, sillimanite schist, and quartzite, as well as rusty pyritiferous quartzite interbedded with actinolite granulite and biotite-quartz schist. The remainder of the rocks, except for a narrow belt of metamorphosed volcanics and quartzite south of South Keene, are intrusives scattered over the area in large and small bodies. The age, sequence, and lithologic character are summarized in a columnar section (Fig. 2). The strata may be traced northward to the Littleton-Moosilauke area where Silurian and Devonian fossils date the rocks (Billings, 1937; Billings and Cleaves, 1934; 1935).

The oldest rocks are the Ammonoosuc volcanics of probable upper Ordovician age. They are 0–600 feet thick and have been metamorphosed to light-colored gneiss and dark amphibolite. Overlying these is a thin sequence of schist not wide enough to map in most places, belonging to the Partridge formation of probable upper Ordovician age. Above these schists lies the Clough quartzite 100 feet or less thick, of lower or middle Silurian age. The Littleton formation of lower Devonian age is

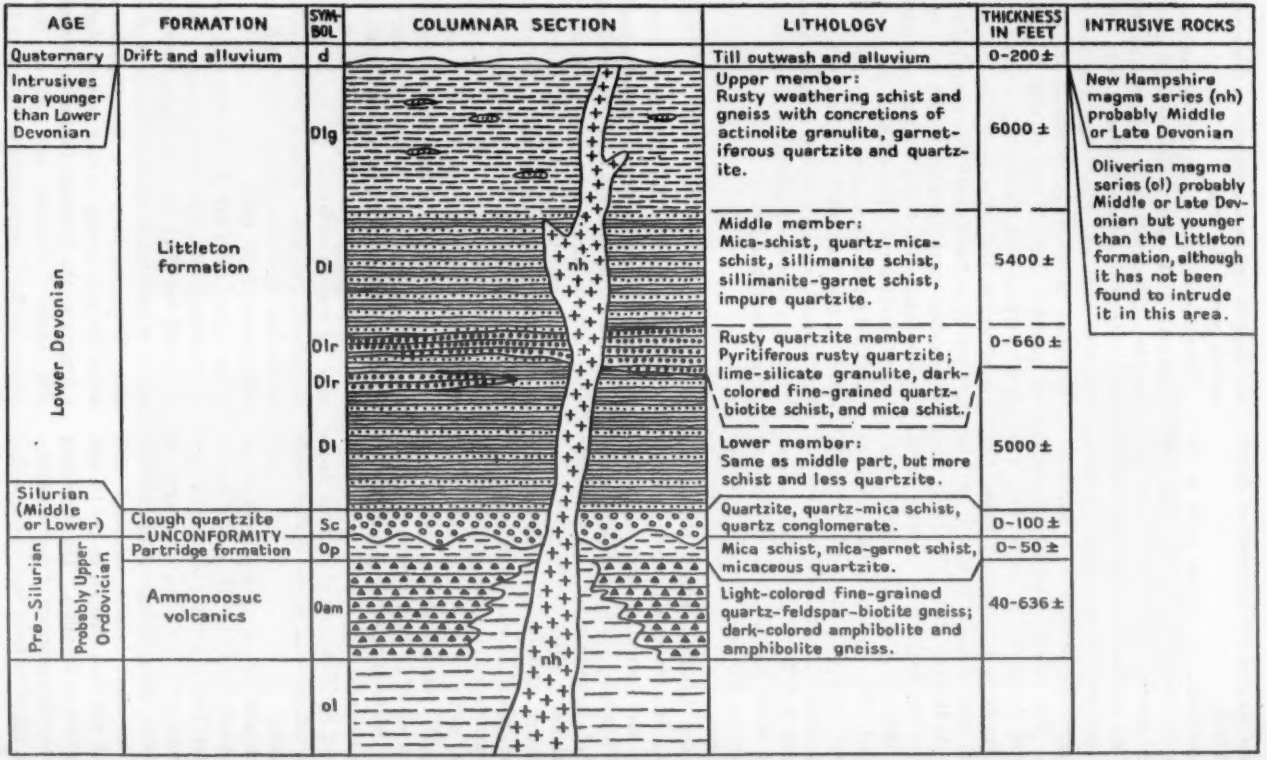


FIGURE 2.—COLUMNAR SECTION FOR MONADNOCK AREA

Shows sequence, character, thickness of the metamorphic rocks and sequence of igneous intrusion. The Oliverian magma series (ol) at the base of the column is not part of the stratified series, but a huge intrusive sheet, the top of which varies considerably in stratigraphic position.

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the youngest of the metamorphosed sediments. It is approximately 17,000 feet thick in this region.

The oldest intrusives found in the western part of the area are quartz diorite and granodiorite belonging to the Oliverian magma series of middle or upper Devonian age. Intrusives of the New Hampshire magma series are next in age; they were intruded during and following the deformation in late Devonian time. Amphibolites are the oldest members of this series. The Kinsman quartz monzonite and the Spaulding quartz diorite were intruded before the cessation of deformation; the Concord granite came in at the close of the folding. Pegmatites are associated with all these intrusives.

Ammonoosuc Volcanics

A narrow belt of Ammonoosuc volcanics 0-600 feet thick is exposed along the western edge of the quadrangle south of South Keene, through Swanzey and Richmond townships. The Ammonoosuc volcanics will be described only briefly here because most of the belt lies west of this area (Fig. 1). G. E. Moore (in press) discusses them more fully. Thin and rather poor sections of the Ammonoosuc volcanics are exposed southeast of Mt. Huggins in the small valley just north of the old lumber road or trail shown on the map; along the railroad cut 1 mile west of Webb; in the gully three-quarters of a mile west of Corey Pond; half a mile south of Falls Brook south of the old road marked on the map; and at the western boundary of the map where the belt of Ammonoosuc volcanics enters the Keene quadrangle 1 mile northwest of Grassy Hill. Several hundred feet of alternating light and dark volcanics may be seen in these localities. Details of five measured sections are given in Table 1.

The Ammonoosuc volcanics consist of light-colored fine-grained quartz-feldspar-biotite gneiss and dark amphibolite and hornblende gneiss. The dark rocks are composed chiefly of hornblende and intermediate plagioclase; the hornblende needles are conspicuous on the foliation surface. Light spots or bands are made up largely of oligoclase-andesine (Table 2, column 1, 2). Beadlike stringers composed of rounded grains of sphene are present. The light-colored rocks in the Ammonoosuc volcanics are fine-grained granular biotite gneiss composed chiefly of feldspar, quartz, and biotite. Garnet occurs in some of these rocks, and actinolite in a few. In many instances, they have been so thoroughly recrystallized that they are difficult to distinguish from fine-grained granite. The feldspar is generally oligoclase-andesine with a little myrmekite and included strained quartz grains (Table 2, column 3). Southeast of Mt. Huggins, a very fine-grained, white fragmental rock in the Ammonoosuc volcanics superficially resembles quartzite, but it is made up of oligoclase (?) and strained quartz; blocks of this same rock were found as inclusions in the granodiorite of the Oliverian magma series near the contact.

The light-colored biotite gneisses are considered to have been soda rhyolite and quartz latite, whereas the amphibolites are considered to have been andesite and basalt. Because the various lithologic types alternate in relatively thin beds a few inches to several feet thick, the volcanics are considered to have been tuffs rather than flows.

Partridge Formation

The Partridge formation is present in several localities overlying the Ammonoosuc volcanics and underlying the Clough quartzite. It is considered to be pre-Silurian, probably upper Ordovician, from correlations with neighboring localities. Since the thickness of the Partridge formation rarely exceeds 50 feet in the Monadnock area, it is shown on the map only at South Keene. Details of its thickness are shown in Table 1, b, c, e. It may be present without outcrop in some of these sections and absent in others.

The Partridge formation is composed of rusty weathering mica schist, mica-garnet schist, sillimanite-garnet schist, and micaceous quartzite. In the railroad cut 1 mile west of Webb, almost 100 feet of rusty weathering schist of the Partridge formation is exposed underlying the Clough quartzite. A thin section shows a granular mica schist, with highly corroded garnet altering to limonite and chlorite (Table 2, column 4). The Partridge

column is not part of the stratified series, but a huge intrusive sheet, the top of which varies considerably in stratigraphic position.

TABLE 1.—*Measured sections of the Ammonoosuc volcanics and Clough quartzite*

		Thickness (feet)
a. Four-tenths of a mile southeast of Mt. Huggins		
Littleton formation	Schists	—
Clough quartzite	White quartzite and quartz-mica-garnet schist	30
Ammonoosuc volcanics	White, very fine-grained quartz-feldspar tuff, resembling quartzite	18
	Light-colored quartz-feldspar gneiss	12
	Black amphibole gneiss	10
	Contact with granodiorite of Oliverian magma series not exposed	
	Total Ammonoosuc volcanics	40
b. Forbush Brook, half a mile southeast of Forbush Hill		
Littleton formation	Mica schists injected by pegmatites	—
	No exposures	50
	Pegmatite	8
Clough quartzite	White quartzite	12
Partridge formation?	No exposures; may represent portion of the Partridge formation	50
Ammonoosuc volcanics	Light-colored quartz-feldspar gneiss	20
	Pegmatite	14
	Black amphibole gneiss	22
	Light-colored quartz-feldspar gneiss	1
	Black amphibole gneiss	8
	Pegmatite	12
	Granodiorite of Oliverian magma series at 700 ft. elevation in Forbush Brook	
	Total Ammonoosuc volcanics	51
c. Railroad section 1 mile west of Webb		
Littleton formation	Rusty weathering, dark-colored biotite schists injected by pegmatite	250
Clough quartzite	Clough quartzite, with some pegmatite at base	50
Partridge formation?	Probable Partridge formation (lower part may be Ammonoosuc volcanics)	110
Ammonoosuc volcanics	Dark-colored gneiss and mica schists containing some actinolite	50
	Mica schist intruded by pegmatite	45
	Black amphibole gneiss	55
	Light-colored quartz-feldspar gneiss	25
	Granodiorite of Oliverian magma series with Ammonoosuc thrust over the granodiorite at contact	
	Total Ammonoosuc volcanics	175
d. Brook section seven-tenths of a mile due west of Corey Pond		
Littleton formation	Mica schist	—
Clough quartzite	Pegmatite, probably marking position of Clough quartzite	10
Ammonoosuc volcanics	Light-colored quartz-feldspar-biotite gneiss	200
	Black mica schist	10
	Light-colored quartz-feldspar-biotite gneiss with some actinolite and garnet	195
	Black amphibole gneiss	105
	Light-colored actinolite gneiss	9
	Black amphibole gneiss	51
	White pyritiferous actinolite gneiss	6
	Black amphibole gneiss	8
	Light-colored quartz-feldspar gneiss	12
	Black amphibole gneiss	40
	Contact with granodiorite of Oliverian magma series at 950 ft. elevation in brook	
	Total Ammonoosuc volcanics	636

TABLE 1.—*Concluded*

		Thickness (feet)
e. One-half mile south of Falls Brook, just south of logging road (see geologic map)		
Littleton formation	Not exposed	—
Clough quartzite?	Pegmatite; marks position of Clough quartzite	100
Partridge formation?	Probable Partridge formation; mica schist	20
Ammonoosuc volcanics	Light-colored quartz-feldspar-biotite gneiss	30
	Not exposed	50
	Black amphibole gneiss	12
	Light-colored quartz-feldspar gneiss	42
	Light-colored actinolite-garnet gneiss	16
	Alternating bands of quartz-feldspar gneiss and amphibole gneiss	33
	Granodiorite of Oliverian magma series; contact at 1110 ft. elevation	
	Total Ammonoosuc volcanics	183

formation is well exposed at the base of the cliff underlying the Clough formation three-quarters of a mile southwest of Brennan Hill.

CLOUGH QUARTZITE

The Clouch quartzite of lower or middle Silurian age is intermittently exposed above the Ammonoosuc volcanics in a narrow belt extending southward from South Keene through Swanzey township to Richmond township. It is characteristically a white or yellowish-white hard vitreous quartzite. Locally there is quartz conglomerate, quartz-muscovite schist, or quartz-garnet-muscovite schist. A thin section of a garnet-muscovite schist shows pink, corroded, fractured garnets, with some sillimanite needles and epidote. The rock is slightly laminated with stringers of fine granular quartz between coarser grains (Table 2, column 5). The Clough quartzite is best exposed half a mile south of South Keene and in the railroad cut 1 mile west of Webb (Table 1, c). South of the railroad cut the Clough quartzite is poorly exposed, and in many places its horizon is represented by pegmatite. It apparently lenses out in many places. South of South Keene, it is at least 100 feet thick. Deformation, irregularities in the original sedimentation, and an unconformity at the top of the formation account for the variable thickness and intermittent exposures of the Clough quartzite in the area.

TABLE 2.—*Approximate modes—Ammonoosuc volcanics, Partridge formation, and Clough quartzite*

	1	2	3	4	5
Quartz	2	20	20	39	89
Potash feldspar			5		
Plagioclase (Oligoclase-andesine)	17	74	68	10	
Biotite			5	22	tr
Muscovite			tr	25	5
Opaques	tr	0.5		3	
Epidote					tr
Hornblende	80	5			
Sphene	0.5				
Garnet			1	tr	5
Apatite				tr	
Sillimanite					tr

Key:

1. Amphibolite, Ammonoosuc volcanics, 0.5 mile southeast of Mt. Huggins (K11)*
2. Hornblende gneiss, Ammonoosuc volcanics, 0.7 mile due west of Corey Pond (K25)
3. Biotite gneiss, Ammonoosuc volcanics, 0.5 mile south of Falls Brook, just south of road (K38)
4. Mica schist, Partridge formation, railroad cut one mile west of Webb (K20)
5. Quartzite, Clough quartzite, 0.5 mile southeast of Mt. Huggins (K14)

* Numbers in parenthesis refer to specimen number.

LITTLETON FORMATION

General Statement

The Littleton formation in the Monadnock area is made up of four members. The lower member is composed of mica schist, quartz-mica schist, sillimanite schist, garnet schist, and impure quartzite, as well as some rocks with retrograde minerals. A rusty quartzite member consisting of pyritiferous rusty-weathering quartzite, actinolite granulite, and dark fine-grained quartz-mica schist makes a con-

spicuous horizon marker in the formation, and has been distinguished on the geologic map (*Dlr*, Pl. 1). The middle member lies above the rusty quartzite and is lithologically similar to the lower member. The lower and middle members have not been distinguished from each other on the geologic map where both are shown by the symbol *Dl* (Pl. 1). The upper member of the Littleton formation (Pl. 1, *Dlg*) consists of rusty-weathering schist and gneiss. It is far more homogeneous than the lower and middle members of the Littleton formation.

Although the Littleton formation in the Monadnock area is in the zone of high-grade metamorphism, some of the rocks contain retrograde minerals typical of lower metamorphic zones.

The total thickness of the Littleton formation in this area is about 17,000 feet. (*See* section on Structure.)

Lower Member of Littleton Formation

The lower member of the Littleton formation, occupying much of the southwestern part of the Monadnock area, lies east of the belt of Ammonoosuc volcanics and Clough quartzite in the western part of the area and southwest of the rusty quartzite member of the Littleton formation. The southeastern part of the area has a heavy cover of glacial drift so that the lower member of the Littleton formation rarely crops out except in Rindge and at the dam in East Jaffrey. It probably underlies the drift south of Bonds Corner. In the township of Marlboro, the lower member of the Littleton formation is confined to the area south of Robbins Brook and west of the village of Marlboro.

An excellent section of the lower member of the Littleton formation can be studied along the highway in the gorge of the South Branch of the Ashuelot River northwest of Troy. It is also well exposed on Little Monadnock Mountain and Grassy Hill.

The rocks of the lower member of the Littleton formation are primarily mica schist, with some garnet-mica schist, impure quartzite, sillimanite schist, and gneiss; graphite flakes are present in many places. Detailed study of the rock types has been limited to the more interesting middle member of the Littleton formation to avoid repetition.

The lower member of the Littleton formation is probably about 5000 feet thick.

Rusty Quartzite Member

The rusty quartzite member of the Littleton formation (Pl. 1, *Dlr*) separates the lower member of the Littleton formation from the middle member. It can be traced fairly continuously from the southeastern end of Thorndike Pond southwest for 2 miles, then north to the Poole Reservoir where it swings southwest along the base of Mt. Monadnock to Gap Mountain. Here it is cut off by a series of intrusives but is picked up on the crest of a hill 1 mile southwest of Gap Mountain and in a single exposure half a mile northeast of Troy village. Northeast along the northwest base of Mt. Monadnock, a belt 4¼ miles long extends to the northeast corner of Dublin Pond, turns west for a mile, then northwest around Hurricane Hill, and is picked up north of the Howe Reservoir where it swings west through the village of Chesham, turns south, and is lost. The horizon normally occupied by the rusty quartzite member swings southwest through the village of Marlboro where it turns northwest to be cut off by the Concord granite north of South Keene.

In the east-central portion of the area two separate beds of the rusty quartzite member are found at the southeast end of Thorndike Pond. These disappear underneath the glacial drift. They probably continue northeast along the strike until cut off by the intrusive Kinsman quartz monzonite.

The rusty quartzite member is composed of pyritiferous rusty-weathering quartzite, actinolite granulite, dark fine-grained granular quartz-biotite schist, and mica schist. It is conspicuous because of the deep red-brown of the quartzite formed by the alteration of pyrite to limonite. The quartzite is massive to thin-bedded and is highly folded in many places. The apparent breadth of outcrop differs greatly from place to place owing to folding.

Where fresh, the rusty-weathering quartzite is dark gray and massive. Microscopically pyrite is conspicuous in stringers parallel to the bedding and makes up 5 per cent of the rock.

TABLE 3—*Approximate modes—Littleton formation*

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26
Quartz	49	28	10	25	29	54	41	40	50	2	5	38	49		42	10	13	25	10	54	24	29	15	19	34	45
Potash feldspar					tr																		35	20	44	
Plagioclase	28	19	24		10	5	10	10	tr			2	5			38	10	36	40	16	12	15		5	5	39
Biotite	15	2	5	18	5		45	22	23	1.5	tr	5	25		10	20		25	30	25	25		24	20	5	
Muscovite							3	22	4		79	39	7	70	20			5				30	tr	3		
Chlorite											tr	10		16	19	20	40					25	tr			
Chloritoid														5	5											
Sericite																	35		tr		35	tr	tr			
Sillimanite								3	20	94	10	tr	5	tr	tr			5	12	2	tr		19	25	5	
Garnet								1			tr	3	5	5				3	5	1	2	tr	5	5	5	9
Opaques				1	tr	tr	tr	1	1	0.5		1	3	3	3		1	tr	2	1	1	tr	1	2	1	0.5
Pyrite	5	tr								1						1										
Tourmaline									1	tr	5	1			tr											
Zircon								tr					tr		tr			tr	tr	tr			tr			
Kyanite											tr															
Graphite								tr																		
Sphene	2		tr	tr	1	tr	tr									5										0.5
Epidote (includes zoisite)			tr	25		30										5										
Actinolite	?	50	60	30	54	10																				5

Key:

Rusty quartzite member of Littleton formation

1. Pyritiferous quartz-plagioclase granulite, north end of Dublin Pond; plagioclase is labradorite (K83)
2. Actinolite granulite, 0.3 mile due west of south end of Dublin Pond (K86)
3. Actinolite granulite, 0.3 mile due west of south end of Dublin Pond (K85)
4. Actinolite granulite, 1 mile southwest of Dublin Pond in brook, 1340 feet altitue (K124)
5. Actinolite granulite, 0.3 mile north of B.M. 1509, Halfway House road (K59)
6. Actinolite granulite, north end of Gilson Pond (K71)
7. Biotite-quartz schist, north end of Farmers Trail, just south of B.M. 1449 (K121)

Middle member of Littleton formation

8. Mica schist, Halfway House (K55)
9. Sillimanite schist, 0.2 mile east of Halfway House (K81)
10. Sillimanite crystal, 0.2 mile east of Halfway House (K81A)
11. Pseudo-sillimanite crystal, Pumpelly Trail, 2960 feet altitude (K99)
12. Muscovite-chlorite schist, Pumpelly Trail, 2960 feet altitude (K98)
13. Sillimanite schist, Gap Mtn., (K47)

Rocks with retrograde minerals in middle member of Littleton formation

14. Muscovite-sillimanite-chlorite-garnet-chloritoid schist, Beech Hill (K91)
15. Mica-chlorite-chloritoid schist, White Arrow Trail, 2825 feet altitude (K53)
16. Chlorite schist, west of Halfway House (K8)
17. Chlorite-sericite schist, 1 mile due south of B.M. 1439, Dublin (K95)

Upper member of Littleton formation

18. Biotite gneiss, north of Chesham Pond (K110)
19. Biotite-garnet-sillimanite gneiss, Marianfield Camp (K109)
20. Biotite gneiss, southeast corner of Skatutakee Lake (K9)
21. Sericite gneiss, south side of Harrisville Pond (K105)
22. Muscovite-chlorite gneiss, south of Harrisville, northwest of Skatutakee Lake (K106)
23. Sillimanite-garnet schist, Cobb Hill (K102)
24. Sillimanite-garnet schist, Cobb Hill (K116A)
25. Gneiss, Cobb Hill (K117)

Concretion

26. Nodular concretion in lower member of Littleton formation, 0.4 mile west of Little Monadnock Mtn. (K40)

Sphene is present as isolated crystals or around the pyrite. Graphite flakes are common. The bulk of the rock is made up of fine granular quartz; labradorite feldspar is present as small grains in some of the beds. Light-brown flakes of biotite are scattered about. The mode of a pyritiferous quartz-plagioclase granulite north of Dublin Pond is given in Table 3, column 1.

The actinolite granulite beds (Table 3, cols. 2, 3, 4, 5, 6) are light to dark greenish gray, granular, slightly laminated, and in places highly contorted. Needles of actinolite are generally aligned in the foliation planes. The granulites contain up to 25 per cent quartz and plagioclase, and up to 50 per cent actinolite. Biotite may be present in small amounts, and

FIGURE 3.—SECTION OF RUSTY QUARTZITE ALONG BROOK 1½ MILES SOUTHWEST OF DUBLIN POND

Section starts at 1340 feet elevation and continues upstream to 1370 feet elevation; 1000 feet of stream section is omitted where stream runs parallel to structure. *1*, highly folded pyritiferous rusty-weathering quartzite projected into section from barn at Weld estate ¼ mile southwest of start of section; *2*, fine-grained biotite-quartz schist, containing concretions and granitic stringers; *3*, well-bedded quartz-biotite schist and impure quartzite; *4*, interbedded mica-sillimanite schist and quartz-mica schist; *5*, mica-sillimanite schist and quartz-biotite schist in small tributary which fills in the gap in section of main stream; *6*, biotite-quartz schist with some mica-sillimanite schist and actinolite granulite layers; *7*, mica schist containing some thinbedded biotite schist, impure quartzite, and mica-sillimanite-garnet schist.

25 per cent epidote or zoisite may occur. Sphene and pyrite are accessories.

The fine-grained, dark biotite-quartz schists and mica schists belonging to the rusty quartzite member are friable and weather easily. They may contain about 40 per cent granular quartz and 40 per cent or more biotite, with small amounts of plagioclase and muscovite (Table 3, column 7).

A section of the rusty quartzite member in a brook on the Weld Estate 1¼ miles southwest of Dublin Pond and three quarters of a mile east of B. M. 1449 at the foot of the Farmers Trail starts near the road at an altitude of 1335 feet and continues upstream to an altitude of 1370 feet (Fig. 3). It is an excellent place to study the stratigraphy and folding of the biotite-quartz schists. The rusty quartzites, exposed by the barn and the house just west of the brook, underlie the biotite-quartz schists, mica schists, and actinolite schists of the brook and are continuous with the section.

Along the road to the Halfway House on Mt. Monadnock, just north of B. M. 1509, at least 40 feet of the rusty, pyritiferous quartzite is overlain by 50 feet of light-colored, friable quartzite and actinolite schist, followed by 30 feet or more of biotite schist. The garnet-sillimanite-biotite schist of the middle part of the Littleton formation overlies the mica schists of the rusty quartzite member. The rusty quartzite is underlain by the Spaulding quartz diorite.

The rusty quartzite member is well exposed at the Poole Reservation southeast of Mt. Monadnock, especially east of the reservoir and at its outlet; on the grounds of the "Ark", an inn 1 mile northwest of the village of Jaffrey; and along the southeast corner of Thorndike Pond where two separate beds of rusty quartzite were mapped. Another interesting section is at the dam at the northern outlet of the Howe Reservoir 1 mile due east of the town of Chesham along the "Eliza Adams gorge" at the outlet of the reservoir. Other excellent exposures can be seen at the northeast end of Dublin Pond; on the land now owned by Mr. Foote a quarter of a mile due west of Dublin Pond; and on the hill 1 mile southwest of Dublin Pond. The rusty quartzites exposed on the grounds of the girls' camp on Derby Hill west of Silver Lake are considered to be a horizon in the upper member of the Littleton formation, for they could not be traced beyond this point.

The thickness of the rusty quartzite member varies, depending upon the amount of folding, from nothing to 600 feet or more.

Middle Member of Littleton Formation

General statement.—Overlying the rusty quartzite member of the Littleton formation are beds of mica schist, sillimanite schist, sillimanite-garnet schist, and impure quartzite which belong to the middle member of the Littleton formation. The rocks are better bedded and contain more quartzite than the lower member of the Littleton formation. Sillimanite and garnet are more abundant than in the lower member. The upper boundary of the middle member extends from Bonds Corner northwest to Skatutakee Lake, west to Chesham Pond, north of Willard, Kidder, and Horse Hills, thence north through Roxbury and Sullivan townships. Mt. Monadnock is made up entirely of the middle member of the Littleton formation and is one of the best exposed continuous sections of the Littleton formation in New Hampshire. Good exposures may also be seen on Beech Hill half a mile northwest of Dublin village and in the gorge of Minnewawa Brook, three-quarters of a mile northwest of the village of Marlboro.

Mica schist.—Mica schist (Table 3, col. 8) is common throughout the middle member of the Littleton formation. It is generally dark, medium-grained with well-oriented biotite and muscovite. Garnet, sillimanite, and flakes of graphite may be present. Granular quartz is the principal mineral; biotite and muscovite appear in equal quantities. Feldspar is subordinate, and a small amount of sillimanite occurs as needles. The biotite is brown and contains zircon with radioactive haloes.

Sillimanite schist and sillimanite-garnet schist.—Sillimanite schist and sillimanite-garnet schist are found in the middle member of the Littleton formation. There is a belt of schist with up to 20 per cent sillimanite on the ridge one-eighth of a mile due east of the Halfway House on Mt. Monadnock (Fowler-Billings, 1944b). The schists are medium-grained dark-brown weathering with large sillimanite crystals in a

groundmass composed of biotite, quartz, and muscovite (Table 3, column 9). A large crystal of sillimanite 5 cm. long and 1.5 cm. wide, from "Point Surprise" on this ridge above the Halfway House, was examined microscopically (Table 3, column 10). Tiny needles of frayed fibrous sillimanite or "fibrolite" have developed in fractures parallel to the basal cleavage as well as parallel to the prismatic cleavage. Quartz penetrates along the fractures as a late replacement. A little brown biotite and magnetite is present in the deformed areas. Tourmaline occurs as small euhedral crystals at the edge of the sillimanite crystal suggesting late introduction by mineralizers from magmatic sources (Pl. 5, figs. 1, 2).

Much of the sillimanite in the schists on Mt. Monadnock has been altered to muscovite by late replacement. Knots or patches of muscovite retain the original outline of large single sillimanite crystals. These resistant "pseudo-sillimanites" show up on the weathered surfaces of the schist slabs (Pl. 4). They lie as a general rule in the plane of schistosity but at different angles to one another; their size depends upon the dimensions of the original sillimanite crystals. Microscopic examination of a pseudo-sillimanite crystal 8 cm. long from Pumpelly Trail, 2960 feet altitude, shows about 80 per cent muscovite and 10 per cent unaltered sillimanite in the core. Small pale to dark-brown tourmalines have developed throughout the pseudo-sillimanite crystal. A little quartz, biotite, and chlorite are present as well as traces of garnet and kyanite (Table 3, col. 11). These pseudomorphs are in a matrix of a dark-brown muscovite-chlorite schist (Table 3, col. 12). Garnet is conspicuous in the matrix of some of the schist beds, so that sillimanite-garnet schists are common.

Large sillimanite crystals have also developed in the schists of the middle part of the Littleton formation at Gap Mountain, 3 miles southwest of the summit of Mt. Monadnock. Blue-gray sillimanite crystals 3 cm. long make up about 20 per cent of a schist bed on the northern knob of Gap Mountain between the 1620- and 1690-foot contours. Microscopically, individual sillimanite crystals 3 cm. long show a fibrous development (fibrolite) at the edges or along cleavage planes of the crystal where there has been strain. The fibrolite radiates in various directions. Up to 1 per cent of tiny magnetite inclusions 0.01 to 0.02 mm. across are present; some of the sillimanite crystals tend to have more inclusions in the center than on the edges. Biotite is present as tiny flakes.

At Gap Mountain, the sillimanite-garnet schist (Table 3, col. 13) has almost been engulfed by the intruding dark Spaulding quartz diorite. Pseudo-sillimanite crystals and schist inclusions are present in the quartz diorite in various stages of assimilation

Impure quartzite.—Massive, gray, impure quartzites are present in many places in the middle member of the Littleton formation. They are beautifully exposed on Mt. Monadnock interbedded with the mica-sillimanite schist (Pl. 3, figs. 3, 4). Individual beds range from a few inches up to 20 feet thick. Although in general graded bedding does not occur, locally the quartzites grade upward into the mica schists and sillimanite schists and aid in determining the top and bottom of the beds. Flakes of biotite and muscovite are scattered through the quartzites; small red garnets are present in some of the beds; and magnetite occurs in small grains. No microscopic study was made of these quartzites; they are similar to those described in the Mt. Washington area (Billings, 1941, p. 892).

Rocks with retrograde minerals.—Some of the rocks of the middle member of the Littleton formation contain chlorite and chloritoid—minerals considered to be retrograde. After the high-grade metamorphic conditions had been attained the temperature fell slowly, and the rocks were maintained for sufficiently long time at lower-grade conditions so that chlorite and chloritoid were developed at the expense of some of the other minerals, especially biotite and perhaps sillimanite.

Beech Hill, 4 miles northeast of Mt. Monadnock, is a locality where 5 per cent chloritoid is present in a muscovite-sillimanite-chlorite-garnet-chloritoid schist (Table 3, col. 14; Pl. 5, fig. 4). The chloritoid gives the rock a bluish-green color. It is retrograde from biotite and appears in thin section as blue lath-shaped crystals. This schist also contains laths of

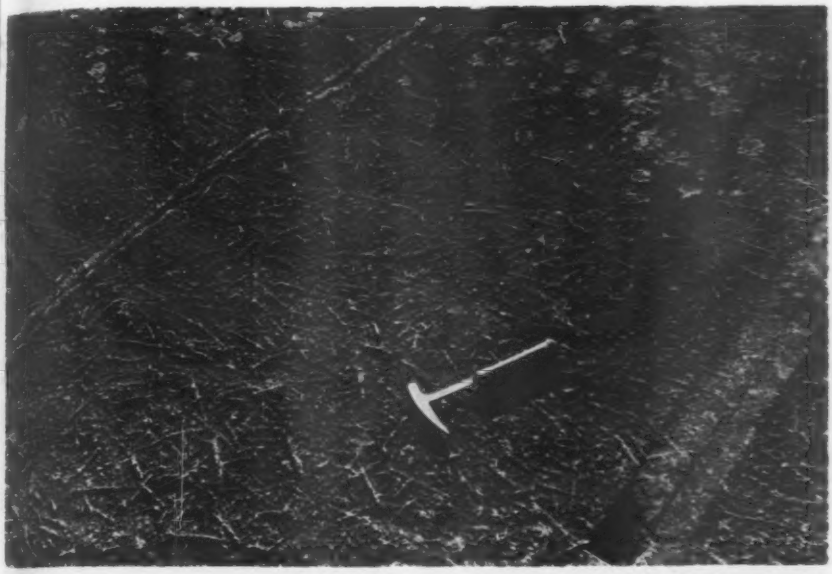


FIGURE 1. PSEUDO-SILLIMANITE CRYSTALS STAND OUT ON WEATHERED MICA SCHIST SURFACE
0.1 mile west of summit of Mt. Monadnock.

FIGURE 2. PSEUDO-SILLIMANITE CRYSTALS
Dark centers are sillimanite; light areas are muscovite; rock is sillimanite-garnet schist, 0.3 mile southeast of summit of Mt. Monadnock.

PSEUDO-SILLIMANITE CRYSTALS

FIGURE 1. PART OF LARGE SINGLE CRYSTAL OF SILLIMANITE
From sillimanite schist of Littleton formation, on ridge ¼ mile due east of Halfway House, Mt. Monadnock. Fibrous sillimanite has formed parallel to prismatic cleavage and in cross fractures.

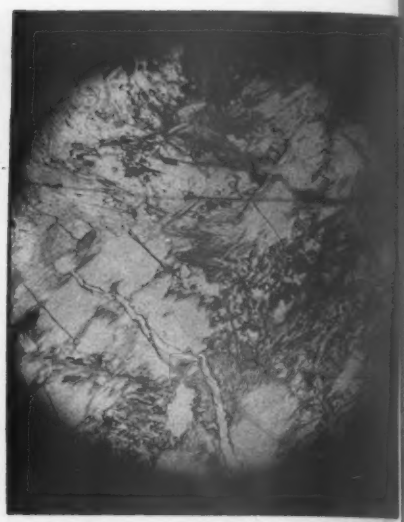


FIGURE 2. SINGLE CRYSTAL OF SILLIMANITE PARTIALLY REPLACED BY FIBROUS SILLIMANITE, TOURMALINE (DARK-GRAY SPOTS), AND QUARTZ
Same locality as figure 1.

FIGURE 3. EYE OF SILLIMANITE
Composed of single large crystal of sillimanite in left half of eye, with small diamond-shaped crystals of sillimanite in right half. Fibrous sillimanite and biotite outline the eye. Littleton formation, Cobb Hill, 1½ miles northeast of village of Harrisville.

FIGURE 4. MUSCOVITE-SILLIMANITE-CHLORITE-GARNET-CHLORITOID SCHIST
From Beech Hill, 4 miles northeast of Mt. Monadnock. Single large crystal of garnet extensively replaced by chlorite. Left third of section is matrix of mica schist composed chiefly of muscovite. Small rectangular grains with high relief are chloritoid.

PHOTOMICROGRAPHS
All × 13 and plane polarized light.

sillimanite 3 cm. long and 0.5 cm. across. The blue-gray sillimanite crystals are pure in many places and altered to muscovite in others. These large crystals are not represented in thin sections so that the mode shows only a trace of sillimanite, although the exposures show 5–10 per cent sillimanite. Deep-red garnets 0.5 cm. in diameter are conspicuous, averaging up to 10 per cent of the schist. The large garnets also were missed in the thin section so that the mode shows only 5 per cent garnet.

Five per cent chloritoid was found in a mica-chlorite-chloritoid schist at an altitude of 2825 feet on the White Arrow Trail on Mt. Monadnock. This schist is composed chiefly of quartz, muscovite, chlorite, and chloritoid with minor amounts of opaques and sillimanite. The quartz is granular and shows strain shadows. The biotite contains magnetite inclusions and small zircon inclusions with prominent radioactive haloes. Some of the muscovite is pseudomorphic after sillimanite. The chlorite appears to be replacing biotite. The chloritoid occurs as deviously oriented crystals up to 2 mm. long (Table 3, col. 15).

A chlorite schist with 20 per cent chlorite and 20 per cent biotite occurs near the Halfway House (Table 3, col. 16).

A chlorite-sericite schist was found 1 mile due south of the village of Dublin along the highway. The schist is fine-grained and compact with highly granular quartz-feldspar aggregates. The high per cent of chlorite and sericite is typical of low-grade zones of metamorphism and represents a retrograde process (Table 3, col. 17).

The middle member of the Littleton formation is about 5400 feet thick.

Upper Member of Littleton Formation

The upper member of the Littleton formation (Pl. 1, *Dlg*) is primarily a rusty-weathering gneiss which occupies the northern part of the Monadnock quadrangle from East Sullivan east to Osgood Hill, and south to Lake Skatutakee. Silver Lake is in the center of the upper member of the Littleton formation. It is well exposed at the southern end of the lake as well as on the grounds of the Marianfield Camp. The Franklin Pierce Highway cuts across a well-exposed section of the gneiss. Most of the hilltops in Roxbury and Nelson townships, notably the summit of Hurd Hill north of Tolman Pond, Parker Hill and Hardy Hill south of Woodward Pond, as well as many others, are good localities to study the upper member of the Littleton formation. The town of Harrisville has typical outcrops of the gneiss in the middle of the village.

At the boundary between the middle and upper members of the Littleton formation, the interbedded quartzites and mica schists grade into structureless, rusty-weathering massive gneisses which overlie them. The line of division is not sharp and is cross-cutting. The upper member, which was primarily shale rather than interbedded sandstone and shale as were the rocks of the middle member, appears to have recrystallized after the end of the deformation, for most of the foliation which may have existed at the close of the folding was destroyed and the gneiss is now relatively massive.

Frost Hill, half a mile southwest of East Sullivan, shows the gneiss typical of the upper part of the Littleton formation. The rock is a granular biotite-quartz-feldspar gneiss which weathers black to rusty brown. In some places the gneiss is covered with a yellow powdery substance which has been derived from pyrite. Graphite flakes are present. The gneiss is highly contorted; elongate nodular concretions of actinolite-garnet granulite and impure quartzite give clues to the original bedding. Small muscovite pegmatites intrude the gneiss.

At the top of Parker Hill, half a mile southwest of Woodward Pond, the upper member of the Littleton formation is well exposed. Here the rusty-weathering, highly granular gneiss contains up to 10 per cent dark-red garnet in grains 1 cm. in diameter. Sillimanite needles 3 mm. long are present. The presence of chlorite indicates retrograde metamorphism.

North of Chesham Pond, dark-gray biotite gneisses are cut by podlike stringers and cross-cutting dikes of muscovite pegmatites. Microscopically, knots of granular quartz and feldspar, 2 cm. in diameter, are conspicuous. The brown biotite is in fan-shaped masses with small amounts of muscovite and shreds of sillimanite associated with it (Table 3, col. 18).

The gray, rusty-weathering biotite-garnet-sillimanite gneiss exposed in the vicinity of Marianfield Camp contains light-colored spots composed of granular quartz and plagioclase feldspar with some microperthite and myrmekite (Table 3, col. 19). Conspicuous rounded, corroded garnets were formed in the early stages. The biotite gneiss at the southeast corner of Silver Lake is similar, but has more quartz and less plagioclase (Table 3, col. 20). Along the south side of Harrisville Pond the gneisses have been sericitized (Table 3, col. 21). The rocks contain large, fractured, corroded garnets, some of which have become augen during the deformation. A muscovite-chlorite gneiss is found south of Harrisville; the chlorite is retrograde (Table 3, col. 22).

At Cobb Hill, 1¼ miles northeast of the village of Harrisville, sillimanite schists carrying up to 20 per cent sillimanite are associated with the gneisses (Fowler-Billings, 1944b). The sillimanite crystals up to 3 cm. long are light gray on the weathered surface of the schist. Each individual crystal is made up of bundles of tiny needles of sillimanite lying parallel to the crystal elongation. Microscopic study shows that single large crystals of sillimanite have developed under pressure into as many as 30 needlelike crystals of sillimanite, or fibrolite, parallel to the elongation of the original large crystal. These bundles of sillimanite show up in diamond-shaped cross section in the basal section of the larger units (Pl. 5, fig. 3). Thus, what appears megascopically to be one individual crystal will show in the basal section 20 to 30 diamond-shaped individuals packed together within the outline of the original single crystal. These compound bundles of sillimanite have become augen from pressure; biotite and frayed and bent needles of fibrolite and muscovite surround the bundles of sillimanite. The sillimanite has a perfect (010) cleavage which makes clear-cut lines across the basal section of the individual diamond-shaped crystals whose faces are (110) prisms. Orthoclase, quartz, a little biotite, and garnet make up the areas outside the eye.

The rock in which these bundles of sillimanite occur is a sillimanite-garnet schist (Table 3, cols. 23, 24). Corroded garnet is intimately mixed with the sillimanite in places. Some of the gneisses near the sillimanite beds are full of small red garnets 0.5 cm. in diameter; pale-green chlorite is present as a retrograde mineral from deep-brown biotite.

Microscopic examination of one of the rusty-weathering, structureless gneisses from Cobb Hill with very little sillimanite reveals augen of strained quartz-feldspar aggregates with fine fibrolite needles bending around the augen. Corroded garnets contain quartz inclusions, as does the microline. The mode of this gneiss which appears to be typical of many of the rocks of the upper member of the Littleton formation is given in Table 3, column 25.

It is estimated that the upper member of the Littleton formation is at least 6000 feet thick.

Nodular Concretions in Littleton Formation

The concretions and elongate quartz-rich masses found commonly in the upper member of the Littleton formation are generally actinolite-garnet granulites. They were originally impure dolomitic concretions or beds. A thin section from one of these compact greenish-gray granulites 1 mile southwest of Little Monadnock Mountain shows large corroded garnets and green actinolite with a little sphene and magnetite in a finely granular mass of quartz and plagioclase (Table 3, col. 26). These concretions also occur in the lower and middle members of the Littleton formation.

Oliverian Magma Series

The eastern boundary of a large plutonic mass enters the western borders of the Monadnock region at South Keene and can be traced south into Richmond township. The rock is a granodiorite or quartz diorite of the Oliverian magma series. It is well exposed on Mt. Huggins. It is medium-grained, equigranular to subporphyritic, granular, slightly foliated, and light gray to pink. The feldspar is chiefly oligoclase-andesine, with some myrmekite and microcline. Strained quartz, biotite, and small amounts of muscovite are present (Table 4, cols. 1, 2, 3).

The correlation of the Oliverian magma series will be discussed with the structure of the east limb of the Swanzey dome, and with the age of the structural features. The larger part of

the intrusive body lies west of the Monadnock quadrangle and is described more fully by G. E. Moore (in press.)

TABLE 4.—*Approximate modes—granodiorite of Oliverian magma series*

	1	2	3
Quartz	20	29	24
Potash feldspar	5	7	5
Plagioclase (Oligoclase-andesine)	65	60	65
Biotite	4	3	5
Muscovite	5		
Pyrite		tr	
Zircon		tr	
Epidote			tr

Key:

1. One-tenth mile due west of South Keene, in brook (K1)
2. Railroad cut 1 mile west of Webb (K21)
3. Three-tenths mile southwest of Beaver Falls, on highway, Keene quadrangle (K61)

NEW HAMPSHIRE MAGMA SERIES

Kinsman Quartz Monzonite

General statement.—The New Hampshire magma series consists of large intrusive bodies which occupy about 40 per cent of the Monadnock area. The oldest of these is the Kinsman quartz monzonite; next in age is the Spaulding quartz diorite; and the youngest is the Concord granite. The Kinsman quartz monzonite occupies an area 7 miles long and 5 miles wide in the northeastern part of the Monadnock quadrangle. It also occurs in smaller bodies near Dublin Pond,[1] East Jaffrey, Sip Pond, Damon Reservoir, south of Fitzwilliam, and elsewhere throughout the area. The Kinsman quartz monzonite can be studied in many accessible places, notably along the shore of Lake Nubanusit, on Skatutakee Mountain, in road cuts southeast of Juggernaut Pond, and on the southeast shore of Sip Pond. The boundaries of the Kinsman quartz monzonite are diagrammatic in most cases due to lack of outcrop.

[1] An error was made in printing Plate 1. The small hill northwest of B.M. 1482 southwest corner of Dublin Pond, should be Kinsman quartz monzonite rather than Littleton formation.

Megascopic description.—The Kinsman quartz monzonite is a coarse-grained, porphyritic, dark-gray quartz monzonite, more rarely a granodiorite or granite, which weathers to a rusty brown. It can be recognized by large phenocrysts of orthoclase which have Carlsbad twinning; microcline and plagioclase phenocrysts are present in a few places. The phenocrysts range in length from 0.5 cm. to 5 cm., and in width from 0.5 cm. to 1.5 cm. Commonly these phenocrysts are aligned and tapered on the ends, for granulation has made many of them augen, with biotite and muscovite bending around the large crystals. Twenty per cent of the rock may be made up of these phenocrysts. In many areas deep-red garnets up to 1 cm. in diameter are peppered through the groundmass; they weather a rusty brown and comprise as much as 10 per cent of the rock.

The groundmass is composed of granular quartz, feldspar, and biotite. The grain size in the coarser varieties averages about 0.5 cm. in diameter. In some places the biotite is roughly aligned and produces a foliated texture; elsewhere, the biotite is not oriented. Blue-green cordierite has been observed in some localities. It is up to 0.5 cm. in diameter, is characteristically altered to chlorite, and is difficult to recognize.

Some of the smaller bodies of Kinsman quartz monzonite, as in the Dublin Pond region, are fine-grained with phenocrysts averaging 0.5 cm. long. In the vicinity of Little Monadnock Mountain, the Kinsman quartz monzonite contains few phenocrysts.

The Kinsman quartz monzonite occurs in the Littleton formation as elongate masses parallel to the bedding. Many of these intrusives are too small to map; others are diagrammatic because of the impossibility of tracing contacts.

Microscopic description.—The essential minerals of the Kinsman quartz monzonite are quartz, potash feldspar, plagioclase, and biotite. Muscovite, garnet, sphene, apatite, zircon, epidote, chlorite, cordierite, and opaque minerals are accessory.

The phenocrysts are generally orthoclase with Carlsbad twinning. Microcline is present as phenocrysts in some sections and may contain

inclusions of quartz, orthoclase, and biotite. Oligoclase-andesine phenocrysts have been found in some thin sections. Some of the phenocrysts are bent and strained, with granulated borders. Myrmekite is present as an end-stage product along the borders of some of the phenocrysts. The myrmekite appears to have developed after deformation for it is undeformed and clear, replacing the finely granulated borders of the microcline phenocrysts.

The groundmass is hypidiomorphic granular and is composed of quartz, oligoclase-andesine, and biotite. The quartz may show strain. The biotite, some of which is oriented, contains zircon inclusions surrounded by radioactive haloes. Some of the biotite contains magnetite inclusions. Garnet (almandite) makes up as much as 10 per cent of some of the slides studied. Many of the garnets are corroded around the borders and altered to biotite in the cracks. Some are altered to chlorite and epidote, especially around the borders, where granulation is intense, suggesting an early development of the garnets. Sphene inclusions were observed in some of the garnets. Cordierite, where present, has been almost completely chloritized.

Approximate modes of the Kinsman quartz monzonite are shown in Table 5. Numbers 1, 2, and 3, from the large body in the northeastern part of the Monadnock area are probably the most typical. Because the rock is coarsely porphyritic, the slides are not always representative. The large body of Kinsman quartz monzonite contains many garnets 0.5 cm. in diameter, but these do not appear in all the thin sections. Some of the smaller bodies are granite, whereas others range from granodiorite to quartz monzonite. Cordierite is difficult to recognize and may be commoner than indicated.

Correlation.—The name Kinsman quartz monzonite has been given to the dark-gray, coarse-grained porphyritic rock of the Monadnock region because of its resemblance to the Kinsman quartz monzonite of other areas to the north—notably the Cardigan region (Fowler-Lunn and Kingsley, 1937, p. 1370); Hitchcock (1877) called it "porphyritic gneiss" and shows it as continuous with the Cardigan body.

Structure indicates that the large body in the northeast part of the region is a batholith. It is the oldest of the large bodies of the New Hampshire magma series in the Monadnock area.

Spaulding Quartz Diorite

General statement.—The Spaulding quartz diorite crops out in several large bodies in the central part of the Monadnock area. The largest of these, in the vicinity of Spaulding Hill and Stone Pond, is 4 miles wide and 5 miles long. There are slightly smaller bodies in the vicinity of Gap Mountain, Cummings Meadow, and west of Laurel Lake, as well as numerous minor elongate masses extending northeast-southwest east of Mt. Monadnock. Some of these are mapped diagrammatically due to the small scale of the map as well as to lack of exposures. Other small ones have been omitted. The most accessible places to see the Spaulding quartz diorite are along the highway on the north side of Cummings Meadow, in road cuts due south of Fassett Brook at the 1340 contour, and on the northern slopes of Gap Mountain around 1500 feet. There are also good exposures along the highway just west of Howe Reservoir and on the summits of Spaulding and Sunset hills. There are small quarries in the Spaulding quartz diorite one-half mile east and 1 mile west of Fitzwilliam Depot.

Megascopic description.—The Spaulding quartz diorite is dark gray and medium-grained. In the larger bodies the rock tends to be massive, whereas in smaller intrusions it is foliated. Weathered surfaces are black to brown. The rock has a distinctive spotted appearance, unlike the other members of the New Hampshire magma series. The dark spots, several millimeters in diameter, are composed of biotite flakes, whereas the light spots, likewise several millimeters in diameter, are composed of granular grains of quartz and feldspar.

. There are places in which inclusions of schist, together with the spotted character of the rock, suggest contamination, but these areas are probably near contact zones, and it is difficult to prove the amount of assimilation. On Gap

TABLE 5.—*Approximate modes—Kinsman quartz monzonite*

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Quartz	14	15	19	20	5	20	5	5	15	19	10	10	15	10	15	20
Potash feldspar	20	15	12	56	76	10	70	54	50	60	49	35	50	35	50	54
Plagioclase (Oligoclase-andesine)	40	44	43	10	10	35	17	25	29	10	20	40	19	42	14	10
Muscovite	tr	8	tr	3		15		3	tr	5	8	4	5	3	5	5
Sericite				5											5	5
Biotite	15	15	15	5	5	19	4	10	5	5	12	10	10	10	10	5
Garnet	10	0.5	10	tr	3	tr	3	tr	tr		tr	tr		tr		
Sphene	tr	tr										tr				
Apatite	tr				tr		tr					tr				
Zircon	tr	tr		tr								tr			tr	tr
Opaques	tr	1		tr	0.1	tr	tr	2	tr		tr	tr	tr	tr	tr	
Pyrite	tr	tr														
Cordierite	tr	tr														
Epidote	tr															
Chlorite	tr	0.5	tr							tr						

Key:

1. On road northeast of Jaquith Brook, 0.2 mile southeast of B.M. 1110; phenocrysts of potash feldspar 5 cm. by 1.5 cm. make up to 20% of rock; garnet grain size is 0.5 cm. (K111)
2. On highway 0.2 mile southeast of Juggernaut Pond; phenocrysts made up of 10% orthoclase and 12% plagioclase are 5 cm. by 1.5 cm.; garnet grain size is 1.5 cm.; biotite, 0.5 cm. (K112)
3. On island in southwest corner of Nubanusit Lake; phenocrysts of potash feldspar and plagioclase 4 cm. long make up 15% of rock; garnet 0.5 cm. in size makes up 10% of rock; biotite is 0.5 cm. average grain size (K104)
4. Halfway between Moose Brook and Hosley Brook, 0.2 mile west of edge of quadrangle on west side of knob, 1300 feet altitude (K113)
5. On cliffs northeast end of Skatutakee Lake (K107)
6. On point northeast end of Skatutakee Lake (K108)
7. On road, west side of Harrisville Pond (K105A)
8. On trial east side of Little Monadnock Mtn., 1430 feet altitude; average grain size 0.5 cm. (K5)
9. On road 0.8 mile north of Beechwood Corners (K41)
10. On route 12, 2.8 miles southeast of Fitzwilliam (K35)
11. On road, 0.4 mile south of Gilmore Pond; phenocrysts of orthoclase 5 cm. long (K64)
12. Highway, 1.6 miles north of East Jaffrey; phenocrysts of potash feldspar 4 cm. long (K70)
13. Halfway between outlet of Gilson Pond and crest of hill 0.1 mile to east; phenocrysts of potash feldspar 1–0.5 cm. long (K72)
14. Two-tenths mile southwest of B.M. 1862, Gap Mtn.; phenocrysts of potash feldspar and plarioclase 2 cm. long (K48)
15. Northeast end of Dublin Pond; phenocrysts of potash feldspar and plagioclase 1 cm. long (K84)
16. South slope of knob, 0.2 mile south of Dublin Pond; phenocrysts of potash feldspar and plagioclase 0.5 cm. long (K93)

Mountain, at about 1500 feet on the northern slopes, pseudosillimanites and partially assimilated inclusions of schist are preserved in the Spaulding quartz diorite. In some of the rock, plagioclase phenocrysts are present as augen; granular quartz and plagioclase form knots at the ends of the crystals; biotite is foliated around the borders.

Microscopic description.—The Spaulding quartz diorite has a typical hypidiomorphic, inequigranular, igneous texture. The essential minerals are quartz, oligoclase-andesine, and biotite, with small amounts of sphene, muscovite, magnetite, ilmenite, and pyrite; there are traces of epidote, zircon, and garnet in some of the slides. Potash feldspar is present in a few areas only. Myrmekite is a late end-stage product. Zoned plagioclase crystals with centers more calcic than the borders, as well as quartz formed around some of the feldspar,

suggest an igneous origin for the rock. Corroded sphene with ilmenite or pyrite cores is typical of most thin sections. Some of the grains of ilmenite are almost entirely replaced by the sphene. The sphene in turn is corroded by the quartz and feldspar. Where potash feldspar is present it may occur as microcline phenocrysts. The biotite is brown to olive green with inclusions of zircon which have radioactive haloes. Epidote, where present, is corroded and encloses other minerals poikilitically. Ten of the modes of the slides studied are quartz diorite, and three are granodiorite (Table 6).

TABLE 6.—*Approximate modes—Spaulding quartz diorite*

	1	2	3	4	5	6	7	8	9	10	11	12	13
Quartz	10	10	10	10	10	5	8	9	10	14	19	9	19
Potash feldspar				5			tr	10	10		tr	tr	
Plagioclase (Oligoclase-andesine)	67	69	58	63	68	52	67	65	64	65	60	60	60
Muscovite (includes sericite)						tr		tr	tr	tr	tr	10	
Biotite	20	19	20	20	19	40	22	15	15	20	20	20	20
Sphene	1	tr	1	1	2	tr	2						tr
Opaques	1	1	tr	tr	tr	tr	tr		tr	tr	tr	tr	tr
Apatite	tr	tr	tr	tr			tr						
Pyrite	tr					tr			tr				
Epidote				tr	0.5	2							
Zircon								tr					
Garnet									tr				

Key:

1. Quarry 0.4 mile east of Fitzwilliam Depot, south of railroad (K32)
2. Quarry 1.4 miles west of Fitzwilliam Depot, north of road (K43)
3. At B.M. 1184, 0.3 mile west of Laurel Lake (K127)
4. On route 124, 0.7 mile southeast of Perkins Pond (K7)
5. From knob 0.6 mile due west of Perkins Pond (K54)
6. Gap Mountain north end, 1500 feet altitude (K90)
7. Highway 0.7 mile south of Meetinghouse Pond (K68)
8. Road 0.6 mile east of Sunset Hill (K77)
9. On route 101, 1 mile west of Howe Reservoir (K88)
10. On road, 2 miles west of Gilmore Pond (K80)
11. Knob, 0.5 mile east of Bonds Corner (K114)
12. On road, 0.3 mile northeast of Rockwood Pond (K4)
13. South of road 1.3 miles south of Cummings Meadow (K66)

Correlation.—Spaulding Hill, in the center of the largest body of the area, has been chosen as the type locality. The Spaulding quartz diorite resembles the Bethlehem gneiss to the north and was given that name in an earlier paper (Fowler-Billings, 1944b). Hitchcock (1877) has previously mapped some of the bodies as "Bethlehem gneiss". The Spaulding is similar to the Hardwick granite of Massachusetts. It is considered younger than the Kinsman quartz monzonite because it is less foliated in most places. In no place have the Kinsman quartz monzonite and the Spaulding quartz diorite been found in contact in this region. Hence their mutual ages cannot be determined directly.

Concord Granite

General statement.—The Concord granite appears as numerous large bodies in the Monadnock area. The largest of these, in the Fitzwilliam area, is 4 miles wide and 8 miles long. A smaller body, 3 miles wide and 2½ miles long, lies due west of Mt. Monadnock. Two smaller areas crop out due north and due south of the village of Marlboro, and a long narrow belt is due north of South Keene. Many smaller

intrusives of Concord granite are scattered throughout the region; some are too small to show on the scale of the map. The boundaries of the intrusives are somewhat diagrammatic, due in part to the lack of outcrop, in part to the penetrating nature of the Concord magma, and in part to the impossibility of showing the interfingering nature of the smaller bodies. The Concord granite has been extensively quarried for building stone; the largest of the quarries are indicated on the map. The Marlboro quarry, 1¼ miles south of the village of Marlboro just east of the highway, as well as the ones on the northern end of Webb Hill half a mile south of Fitzwilliam Depot, are readily accessible. The jointing of the Concord granite, the variations in the granite, and the small pegmatites cutting it are clearly exposed in the quarries (Pl. 3, fig. 2).

Megascopic description.—The Concord granite is a light-gray binary granite. It is fine- to medium-grained, locally porphyritic, and slightly foliated in many exposures. Pegmatites, associated with the granite as late end-stage phenomena, range in width from a fraction of an inch to several feet.

Microscopic description.—The Concord granite is a hypidiomorphic granular igneous rock; the essential minerals are quartz, potash feldspar, plagioclase (oligoclase-andesine), biotite, and muscovite, with such accessories as apatite sphene, zircon, and garnet. Garnet is visible megascopically in many places, but was not present in the slides studied. Many of the slides show areas of fine-grained granular quartz and feldspar with some end-stage myrmekite and perthite surrounding larger grains of microcline and orthoclase. Locally, the rock contains phenocrysts of potash feldspar up to 1 cm. in size. The muscovite and biotite are poorly aligned in many of the slides.

The average of the modes given in Table 7 is granite, but there are specimens which can be called quartz monzonite or even granodiorite.

Correlation.—The name "Concord granite" is applied to the light-gray, binary, fine- to

TABLE 7.—*Approximate modes—Concord granite*

	1	2	3	4	5	6	7	8	9	10	11	12	13
Quartz	18	10	15	15	10	20	19	20	19	15	10	14	20
Microcline (includes orthoclase)	56	50	25	34	34	45	20	70	57	54	15	53	69
Plagioclase (Oligoclase-andesine)	15	25	40	30	50	29	35			20	53	10	
Muscovite	5	8	10	10	5		10	6.5	3	5	20	20	7
Biotite	5	6	9	10		5	15	2.5	20	5	1	2	3
Chlorite					1								
Apatite	tr	tr											
Sphene						1			tr				
Opaques		tr							tr				
Zircon									tr				

Key:

1. Marlboro quarry, east of route 124, 1.2 miles south of Marlboro (K29)
2. Quarry 0.2 mile south of route 119, 0.6 mile east-southeast of Fitzwilliam (K42)
3. Quarry on knob 0.7 mile north of South Keene (K24)
4. 0.7 mile northeast of Mt. Huggins (K23)
5. 0.6 mile east-northeast of Mt. Huggins (K17)
6. 0.4 mile northeast of Mt. Huggins (K22)
7. Power line, 1 mile east-northeast of Mt. Huggins (K16)
8. At west edge of Monadnock quadrangle on Franklin Pierce Highway (K6)
9. 0.8 mile west-southwest of Mt. Monadnock in brook, 1850 feet altitude (K79)
10. 0.2 mile north of B.M. 1862, Gap Mtn. (K49)
11. On road 0.1 mile northeast of Fulham Hill (K52)
12. 0.4 mile southeast of West Rindge, south of route 202 (K69)
13. Due west of Lake Monomonac on State Line on highway (K118)

medium-grained, massive to slightly foliated granites of the Monadnock region, following the terminology of Hitchcock (1877, vol. II, p. 24 and atlas). The Concord granite closely resembles similar light-gray, fine-grained granites in the Cardigan area (Fowler-Lunn and Kingsley, 1937, p. 1371). A similar granite in the Franconia and Mt. Washington area (Billings, 1941) has been called the "Bickford granite". Emerson (1917, p. 238) called this granite "Fitzwilliam granite" comparing it to the Fitchburg granite to the south. The Concord granite is the youngest of the New Hampshire magma series, for it cuts the Kinsman quartz monzonite and the Spaulding quartz diorite.

Dikes and Sills

Amphibolite Dikes and Sills

There are three kinds of dikes in the Monadnock region: amphibolite, pegmatite, and biotite schists. No unmetamorphosed mafic dikes were found.

Amphibolite dikes and sills intrude the Oliverian magma series 4½ miles south of South Keene, half a mile west of the contact of the Oliverian magma series with the Ammonoosuc volcanics. Amphibolite dikes and sills also cut the Ammonoosuc volcanics and, in a few instances, the Littleton formation. They have not been observed to cut the Kinsman quartz monzonite, the Spaulding quartz diorite, or the Concord granite. They were originally mafic dikes and sills older than the regional metamorphism. Since they intrude the Oliverian magma series, the latter must likewise be older than the metamorphism. They are considered to be the earliest members of the New Hampshire magma series. G. E. Moore (in press) discusses the seintrusives more fully in his paper on the Keene-Brattleboro region where they are more abundant.

Pegmatites and Quartz Veins

Pegmatites are abundant throughout the Monadnock region, associated with the late stages of the Oliverian and New Hampshire magma series. The largest pegmatites are in the northwestern part of the area in Sullivan, Roxbury, and Marlboro (Pl. 1). They vary in thickness from a fraction of an inch to [illegible] feet. Most of them are unfoliated. Some are parallel to the structure of the adjacent rocks, others cut across it. Potash feldspar, quartz, muscovite, biotite, black tourmaline, and garnet are the usual minerals in these pegmatites. Most of them belong to the New Hampshire magma series and were intruded after deformation, in the latter part of Devonian time. Each magma series had its pegmatitic juices associated with individual intrusions. It is not always possible to determine from which of the intrusives the pegmatites were derived. In general, the pegmatites in the schists came from the nearest intrusive body. Pegmatites presumably derived from the Oliverian magma series and intruding the Ammonoosuc volcanics are intensely folded, which proves their intrusion before the deformation of the region. Some of the small pegmatites in the Littleton formation are slightly deformed, indicating intrusion before the folding was complete. They came, presumably, from the Kinsman quartz monzonite or the Spaulding quartz diorite which were intruded before folding had been completed. Since the Concord granite was intruded at the close of deformation, the pegmatites associated with this rock are not deformed.

Graphite occurs occasionally in shear zones or associated with small quartz veins. It was mined locally in four localities about 100 years ago (Pl. 1). In one of these abandoned pits, 1¼ miles southwest of the village of Nelson, the graphite occurs in quartz veins and pods of pegmatite, but very little graphite was found. Some specular graphite was found sparsely disseminated in a rusty quartzitic bed in the gneiss of the upper member of the Littleton formation at this locality. Another prospect on Mt. Monadnock, just south of Bald Knob at an altitude of 2560 feet, half a mile northeast of the Halfway House, contains a dull granular type of graphite in shear zones in a cross-cutting quartz vein.

An abandoned gold prospect 1 mile due east of School No. 3 in Dublin shows pyrite in quartz veins. It is questionable whether much gold was taken from the pit.

Biotite Schist Dike

A dike composed of fine-grained, dark biotite schist, weathering to a rusty brown, can be seen east of the summit of Mt. Monadnock on the Red Cross Trail at 2940-foot altitude. The dike cuts across the structure, striking N.25°E.; it is vertical or dips steeply west. Microscopically, the dike is a fine-grained, granular biotite-quartz-plagioclase schist, with at least 50 per cent biotite, 33 per cent quartz, 15 per cent plagioclase, and a little pyrite. It is obviously not of igneous origin and resembles some of the metasedimentary dikes of the Mt. Washington area (Fowler-Billings, 1944a). During metamorphism, plastic sedimentary material must have been forced into an opening in the less plastic rocks above forming a dikelike mass. Since the dike is a biotite schist cross-cutting the structure, it must have been intruded after the rocks were folded, but before metamorphism was completed.

Structure

General Statement

The western part of the Monadnock area is dominated by the east limb of a large anticline. The rocks in the central part of the region are complexly folded schists and gneisses of the Littleton formation with northeast strikes and steep dips. Mt. Monadnock is a large syncline in the middle of the folded Littleton formation. The manner in which the rusty quartzite member of the Littleton formation zigzags back and forth across the area indicates a series of anticlines and synclines plunging northeast. Minor folds are characteristic of the folded sediments.

Large igneous bodies are prominent in the area. The foliated rocks of the Oliverian magma series are part of a large concordant intrusion exposed in the east limb of the anticline south of South Keene. The New Hampshire magma series is represented by a large foliated discordant body of Kinsman quartz monzonite in the northeastern corner of the Monadnock quadrangle. Small, elongate, foliated, northeasterly striking sills of the Kinsman quartz monzonite are found southwest of the main body as far south as the Massachusetts State line. The less foliated Spaulding quartz diorite appears in the core of an anticline in the center of the Monadnock quadrangle. Discordant and concordant smaller bodies of the Spaulding quartz diorite are present in the south-central part of the area. The Concord granite is the least foliated of the New Hampshire magma series. It is largely cross-cutting and appears in a large body in the south-central section, as well as in smaller bodies west of Mt. Monadnock, north and south of the village of Marlboro, and north of South Keene.

Faults are represented by four separate northeasterly striking silicified zones in the south-central part of the area.

Three structure sections accompany the geologic map, on Plate 1. The structure map (Pl. 6) should be consulted for structural details.

East Limb of Swanzey Dome

The east limb of a large anticline, called the "Swanzey dome" in this article, is exposed in a strip 3 miles wide along the western border of the Monadnock area. Since the larger part of the dome lies west of the Monadnock area, it is described more fully by G. E Moore (in press).

The core of the anticline is a concordant body of granodiorite of the Oliverian magma series. The strike of the foliation of the granodiorite swings from an average of N. 20°W. in the north near South Keene, to N. in the vicinity of the railroad, to N.45°E. south of Bridge Brook. The dips are from 30° to 50° E. The foliation, which is more prominent near the contact, is in part primary and in part due to later deformation (Hadley, 1942, p. 163). A secondary granulation with strained quartz is characteristic of the granodiorite.

The foliation and bedding of the Ammonoosuc volcanics which overlie the granodiorite of the Oliverian magma series are parallel to the foliation of the intrusive. The contact plane is presumably parallel to the foliation and dips about 40° E. Hadley (1942, p. 161) and Moore (in press) have observed this relationship in the Mt. Cube and Keene-Brattleboro areas. The Oliverian magma series is intruded into the upper part of the

Ammonoosuc volcanics in the Monadnock region.

The only contact observed in the Monadnock area was a thrust fault, along the railroad cut 1 mile west of Webb (Fig. 4). The Ammonoosuc volcanics were thrust over the grano-

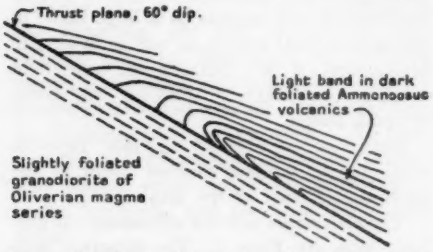

Thrust plane, 60° dip.
Light band in dark foliated Ammonoosuc volcanics
Slightly foliated granodiorite of Oliverian magma series

FIGURE 4.—THRUST CONTACT OF GRANODIORITE OF OLIVERIAN MAGMA SERIES AND AMMONOOSUC VOLCANICS

North wall of railroad cut 1 mile west of Webb. Due to perspective, the apparent dip of the thrust plane is considerably less than the true dip, 60°.

diorite from the east. The amount of the net slip could not be determined. The thrust plane strikes N. 5°E. and dips 60°E. Slickensides on the fault plane plunge 58°E. The foliation of the granodiorite is parallel to the thrust plane. The strike of the foliation and bedding of the Ammonoosuc volcanics is the same as that of the fault, but the dip is lower. A thin light-colored bed in the Ammonoosuc volcanics shows that this formation is folded. The axial plane of the fold strikes N. and dips 45°E.; the axis plunges 20°N.40°E. The variable breadth of outcrop of the Ammonoosuc volcanics along the east limb of the anticline may be due to this type of folding and thrusting as well as to variations in the upper limit of intrusion of the Oliverian magma series.

Evidence elsewhere also shows that the Ammonoosuc volcanics have been folded. For example, minor folds are present in a section three-quarters of a mile west of Corey Pond. The bedding in general strikes N. and dips 50°E.; the axial planes of the minor folds strike NW. and dip steeply NE.; the axes plunge 35°N. The lineation shown by actinolite needles in one of the light-colored beds parallels the axes of the folds.

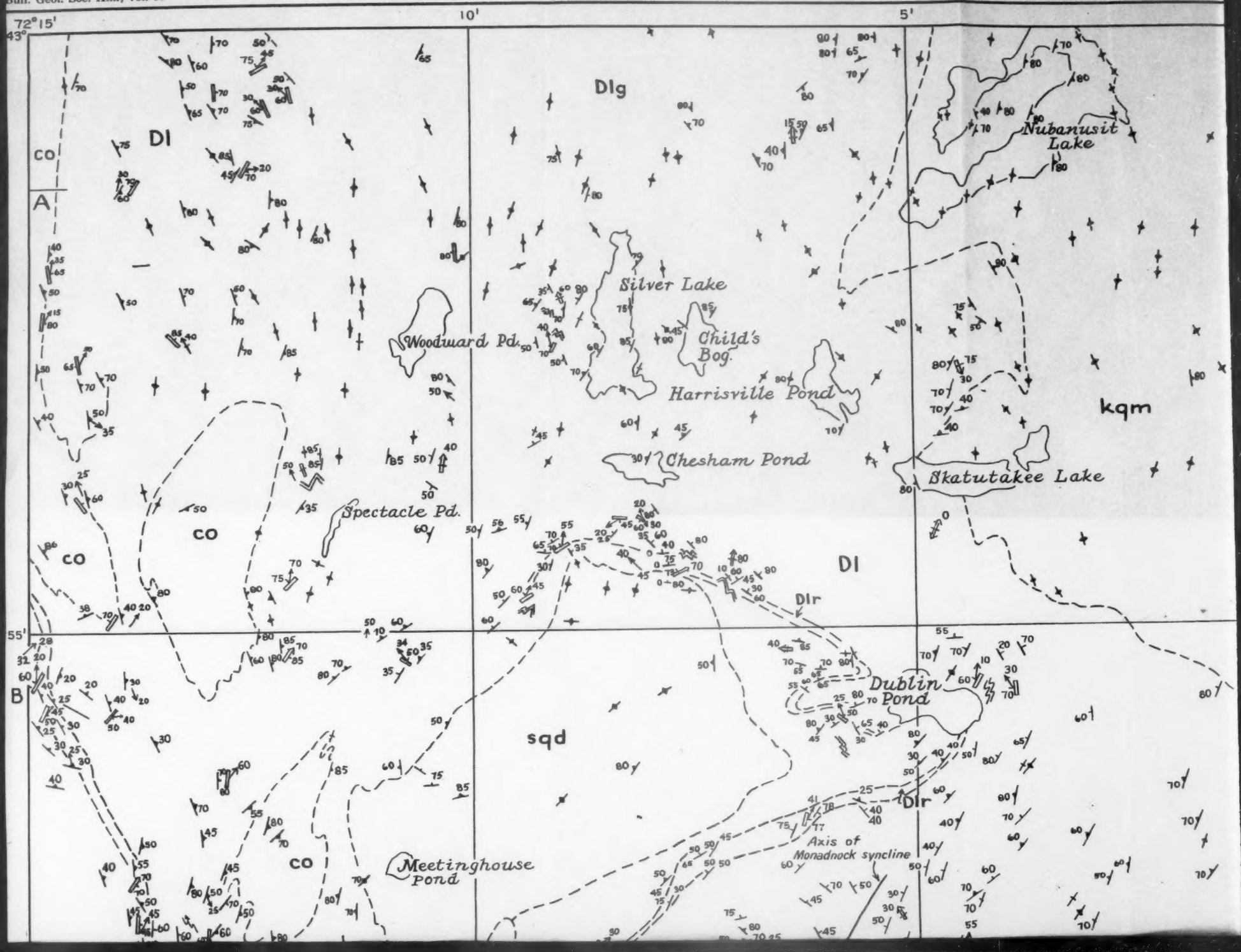
The Partridge formation, where present above the Ammonoosuc volcanics, has the same attitude as the underlying rocks. It is highly schistose and shows an enormous amount of crinkling and minor folding. Half a mile southwest of Brennan Hill, 50 feet of schist of the Partridge formation is exposed at the base of the Clough quartzite. The schistosity strikes N.20°E. and dips 40°E.; the axial planes of small crinkles strike NE. and dip steeply SE.; the axes of the crinkles plunge 20°NE.

The intermittent exposures of Clough quartzite above the Ammonoosuc volcanics and the Partridge formation have the same attitude as the underlying rocks. Bedding, schistosity, and folding are present in many of the outcrops; in others, the quartzite is massive. In the section half a mile due south of South Keene, the Clough quartzite is highly folded; the axial planes strike NE. and dip 60°NW.; the axes strike NE. and plunge 20°N. The bedding and schistosity are parallel and strike N.10°W. and dip 40°E. Southeast of Mt. Huggins the Clough quartzite strikes N.90°E. and dips north; a lineation dips 25°N. on the bedding planes.

The Littleton formation directly overlying the Clough quartzite is foliated parallel to the underlying rocks. The entire 3-mile wide belt of Littleton formation in the western part of the Monadnock area is a part of the east limb of the Swanzey dome. The strike of bedding and foliation is N. in general, with steep easterly dips. The rocks appear to be more or less isoclinally folded. The axes of the minor folds plunge northeast; variable northeasterly dips in the Littleton formation are present in the area east of the Clough quartzite contact and near the village of Marlboro. In the vicinity of Little Monadnock Mountain and throughout the southwestern part of the area, the axes of the minor folds tend to have a southeast strike with southerly plunge. The lineation parallels the axes of the folds and plunges 20° south.

Folded Rocks of Littleton Formation

Mt. Monadnock syncline.—The rocks in the central part of the Monadnock region are complexly folded gneisses and schists of the Littleton formation. The best exposures to show the lithologic character and structure of


72°15'
10'
5'
43°
55'
A
B
Dlg
Dl
Dlr
sqd
kqm
co
Nubanusit Lake
Silver Lake
Woodward Pd.
Child's Bog
Harrisville Pond
Chesham Pond
Skatutakee Lake
Spectacle Pd.
Dublin Pond
Meetinghouse Pond
Axis of Monadnock syncline

72°15' 10' 5' 72°

43° 43°

55' 55'

Dl

Dlg

co

kqm

sqd

Dlr

Nubanusit Lake

Silver Lake

Woodward Pd.

Child's Bog

Harrisville Pond

Chesham Pond

Skatutakee Lake

Spectacle Pd.

Dublin Pond

Meetinghouse Pond

Axis of Monadnock syncline

A A'

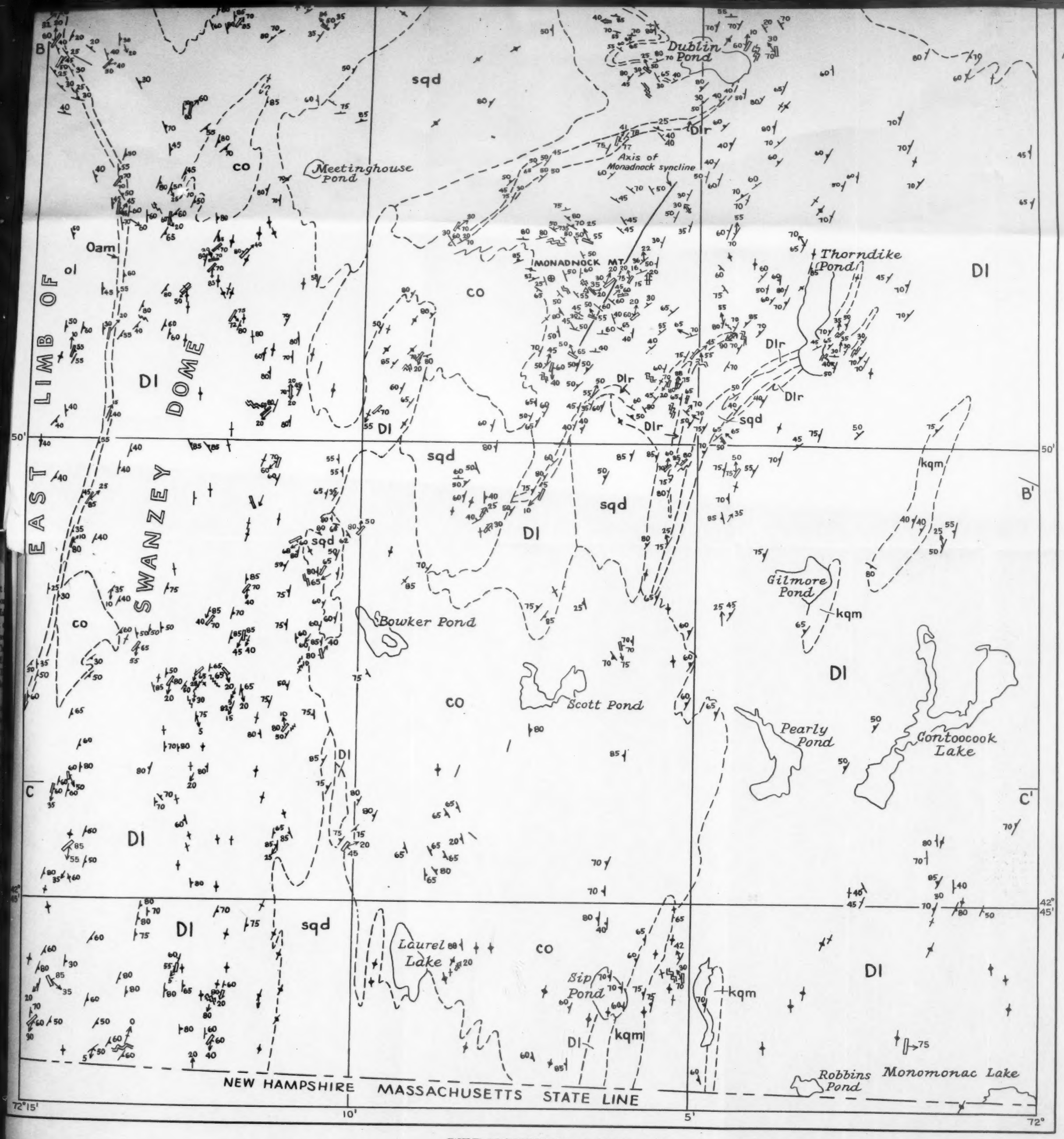
B

LEGEND

- 30 Bedding
- ⊕ Bedding where horizontal
- × Bedding where vertical
- 30/40 Bedding and lineation
- 50 Foliation (schistosity)
- Foliation where vertical
- Foliation with undetermined dip.
- 30/50 Foliation (schistosity) and lineation.
- 20/50 Fold – double line gives strike of axial plane, and figure gives dip; arrow gives strike of horizontal projection of axis with value of plunge.
- 25 Fold – vertical axial plane. Figure gives value of plunge of fold.
- 25 Lineation
- Plan of folds
- — — Approximate boundaries of formations.
- A-A' Ends of structure sections.

FORMATION SYMBOLS

- co Concord granite
- sqd Spaulding quartz diorite.
- kqm Kinsman quartz monzonite.
- ol Oliverian magma series
- Dlg Upper part of Littleton formation.



— — Approximate boundaries of formations.

A–A' Ends of structure sections.

FORMATION SYMBOLS

co	Concord granite
sqd	Spaulding quartz diorite.
kqm	Kinsman quartz monzonite.
ol	Oliverian magma series
Dlg	Upper part of Littleton formation.
Dlr	Rusty quartzite of Littleton formation.
Dl	Littleton formation
Oam	Ammonoosuc volcanics, Partridge formation, and Clough quartzite.

0 — 1 — 2
Scale in Miles

STRUCTURE MAP

Shows structural data by appropriate symbols. Only part of the field observations could be recorded on this scale.

the Littleton formation in New Hampshire are found on Mt. Monadnock.

Mt. Monadnock is a large syncline (Pl. 6) plunging northeast. Structure section BB′ (Pl. 1) cuts across the summit of the mountain in a northwesterly direction and transects the 3-mile wide syncline. Only a few of the many readings taken in traverses across the region could be plotted on the scale of the map (Pl. 6). In general, the beds on the east side of the mountain, between the rusty quartzite and a point a quarter of a mile east of the summit, dip NW. and constitute the east limb of the syncline. The axis of the syncline can be seen half a mile east of the summit along the Pumpelly Trail at 2800 feet. A large rectangular glacial erratic of schist of the Littleton formation, locally known as the "Sarcophagus", lies right in the axis where the beds swing from a NE. strike and 50° NW. dip, through the E.–W. direction with northerly dips, to a NW. strike and 40°E. dips. (Chamberlin's map, 1940, gives trails and local points not shown on the U. S. topographic map.) Pumpelly Ridge, which the Pumpelly Trail follows, is held up by the southeastern limb of the syncline; the Farmers Trail follows the northwest limb of the syncline; and a north-flowing brook lies about in the axis. The summit of Mt. Monadnock lies slightly to the west of the axis of the syncline and is held up by resistant quartzites and pseudo-sillimanite schists which have steep to 40°E. dips. A small drag fold with a 20° NE. plunge complicates the structure at the summit and makes a broad arc with strikes swinging from NE. to E.–W., and then to NW. (Pl. 6). Southeast of the summit along the Red Cross Trail, the rocks have been thrown into minor folds, so that the trail follows along approximately the same bed of quartzite on the west limb of the syncline for almost a quarter of a mile.

Exposures northwest of the summit of Mt. Monadnock along the Farmers Trail above the 1900-foot contour show NW. strikes and steep NE. dips. The trail follows the northwest limb of the syncline (structure section BB′, Pl. 1).

The northwest limb of the syncline is beautifully exposed on the Marlboro Trail due west of the summit of Mt. Monadnock (Pl. 1). The strikes are N. and the dips 50°E. Three-quarters of a mile due west of the summit along the Marlboro Trail, the rocks are horizontal. This point marks the change from the easterly dipping beds of the NW. limb of the Monadnock syncline to westerly dipping beds in a minor anticlinal drag fold.

Due west of the summit of Mt. Monadnock, on a south-facing cliff west of the White Arrow Trail about 100 feet vertically below the summit, is a conspicuous drag fold 18 feet high and 30 feet long in the quartzites and pseudo-sillimanite schists (Pl. 3, fig. 4). The fold is an overturned anticline, the axial plane of which strikes NW. and dips 30° NE. The axis plunges 20° NE. The fold has been cut at an angle that makes it appear to be recumbent. Since the strike of its axis and its axial plane is not in keeping with the general structure of the top of the mountain, this overturned fold can be interpreted as having been formed while the sediments were still soft. Other minor drag folds, also overturned, are found in the same approximate horizon at lower altitudes near the White Arrow Trail. Their axial planes strike NW. and dip NE., and the axes plunge NE.

Lineation of the biotite or sillimanite crystals is present in some places on the foliation planes or on the bedding. The readings have been plotted on the structural map (Pl. 6) where possible. In general, the lineation on Mt. Monadnock plunges 45° NW., N., or NE.

The joints on Mt. Monadnock are conspicuous (Pl. 3, Fig. 1). There are at least three sets at high angles to each other; they cut across the major structures and therefore help little in interpreting the general structure of the syncline. A prominent northeast set of joints dips vertically to steep southeast; a set at almost right angles to these has southwesterly dips; east-west joints dip steeply north or south; and low-angle joints have developed more or less parallel to the surface as sheeting due to relief in pressure upon uplift and erosion. The steep cliffs east of the Pumpelly Ridge Trail are due to glacial plucking in the joints of the northwesterly dipping schists and quartzites. Perry (1904, p. 10) lists the strike and dip of the major sets of joints of Mt. Monadnock.

Veins a quarter of an inch or more thick and composed of very fine black tourmaline are conspicuous in many places near the summit of

Mt. Monadnock. They occupy shear zones which cut across the regional structure at various angles.

Evidence of rusty quartzite member.—The rusty quartzite member of the Littleton formation gives the main clue to the structure of the central part of the Monadnock area. It zigzags across the central part of the region outlining the position of anticlines and synclines (Pl. 1).

East of Thorndike Pond are two belts of rusty quartzite, highly folded with NE. strike and 40°–70° NW. dips. The minor folds plunge northeasterly. The exposures at the "Ark", 1¼ miles southwest of Thorndike Pond, are a part of this same limb. To the southwest, this limb is cut off by the Spaulding quartz diorite. The more easterly bed of rusty quartzite presumably thins out (Pl. 1, section BB').

The western limb of this syncline appears east of Cummings Meadow, strikes north, and is overturned with vertical to steep westerly dips. It may be followed north beyond the Poole Reservoir, which lies between the two limbs of an anticline just south of the nose of the fold. The western limb strikes northeast and dips northwest underneath the Monadnock syncline. The rusty quartzite in the nose of the anticline can be followed for almost a mile north of the Poole Reservoir where it disappears in the vicinity of the Harling Trail. The northwest limb swings west in a broad arc around the base of Mt. Monadnock dipping 45°–75° NW. underneath the mountain. The actinolite granulite and biotite gneiss on the east shore of Gilson Pond (colored pink but not labeled on Pl. 1) may represent a reappearance of the rusty quartzite member in a double-plunging minor fold; or they may be higher beds in the Littleton formation.

South of the Jaffrey-Marlboro highway the rusty quartzite on the northwest limb of the Poole Reservoir anticline is discontinuously exposed. Exposures of rusty quartzite on this limb appear on Gap Mountain and 1 mile to the southwest. Presumably the exposures of rusty quartzite 1 mile southwest of Gap Mountain and a half mile northeast of the village of Troy were continuous before the intrusion of the Concord granite. Likewise the exposures northeast of Troy probably are an extension of the Gleason Brook belt of the rusty quartzite member 3½ miles northeast of Troy. A large body of Concord granite has cut out the belt in this area.

The 5-mile long belt of rusty quartzite which strikes northeast from Gleason Brook to the eastern end of Dublin Pond dips 50° or more to the northwest. It lies on the overturned northwestern limb of the Monadnock syncline. A detailed section of the rusty quartzite member along the brook between the altitudes of 1340 and 1370 feet, 1½ miles southwest of Dublin Pond, is given in Figure 3. The rusty quartzite is well exposed by the house and barn a few hundred yards west of the brook section. The rocks strike N. 20° E. and dip between 65° W. and vertical. The drag folds plunge northeast at an average angle of 30° to 40°; the axial planes dip southeast. Some of the drag folds indicate that the beds to the southeast sheared upward relative to the beds to the northwest consistent with the concept of a major syncline to the southeast. Other drag folds, however, suggest that the rocks to the northwest sheared upward relative to the rocks to the southeast, the reverse of what one would expect on the overturned northwesterly limb of a syncline (Fig. 5). The reversal of some of the drags here and elsewhere along this overturned limb could be explained by the forceful intrusion of the body of Spaulding quartz diorite to the northwest.

The rusty quartzite member swings west from Dublin Pond and then northwest from Hurricane Hill in a series of minor folds which are structurally the eastern nose of the broad anticline which wraps around the large intrusive body of Spaulding quartz diorite. The pattern of the minor folds in and around Dublin Pond is than of an anticline plunging NE. Many folds, however, plunge NW. The same is true of the folds in the vicinity of the village of Chesham which lies on the north limb of this anticline. The forceful injection of the Spaulding quartz diorite is presumably responsible for the NW. plunges.

It will be recalled that the rusty quartzite on the NW. limb of the anticline disappears south of Chesham, and that the horizon in which it presumably lies goes southwest through the village of Marlboro, thence northwest toward the border of the map. This implies another large syncline to the north of Marlboro.

Rest of the Littleton formation.—The structure of the upper member of the Littleton formation in the area north of Horse Hill, Chesham Pond, and Skatutakee Lake is difficult to determine because of the lack of bedding. Where schistosity or bedding are present, N.-S. to NE. strikes and steep dips are the rule. A series of minor folds which tend to be open rather than isoclinal (Pl. 1, structure section AA′) are suggested by the folds of a rusty quartzite bed at Derby Hill. This bed is a rusty quartzite in the upper member of the Littleton formation, and was not traced beyond here. The two isolated exposures of rusty quartzite in the extreme northeast corner of the Monadnock quadrangle (Pl. 1, Dlr) are roof pendants in the Kinsman quartz monzonite. They strike NE. and dip steeply NW. There is no way of proving to which member of the Littleton formation they belong.

The Littleton formation in the south and southeastern part of the Monadnock region is considered to be closely folded (Pl. 1, structure section CC′). Lack of key horizons, the large intrusive bodies, and a thick covering of glacial drift make it impossible to obtain many data on the structure. The strikes are NE., and the dips are steep to northwest for the most part.

Thickness of Littleton Formation

The lower member of the Littleton formition from the top of the Clough quartzite to the base of the rusty quartzite member is about 5000 feet. This figure is based on the distance, 1½ miles, from a point southeast of Mt. Huggins to the village of Marlboro, the shortest distance between the Clough quartzite and the presumed top of the lower member of the Littleton formation. Allowing for some isoclinal folding, 5000 feet seems a fair approximation of its thickness.

The rusty quartzite member is highly folded with thickening and thinning in many places. Where least folded its total thickness is probably not much over 600 feet. Duplication due to folding has increased its apparent thickness two or three times in places. Elsewhere it has been thinned tectonically so that it is only a few feet thick, or it has been squeezed out.

The middle member of the Littleton formation is in the neighborhood of 5400 feet thick. This figure is based on the section on Mt. Monadnock which probably represents almost the entire thickness of the middle member (Pl. 1, structure section BB′). Probably erosion has removed some of the middle member of the formation from the top of the syncline.

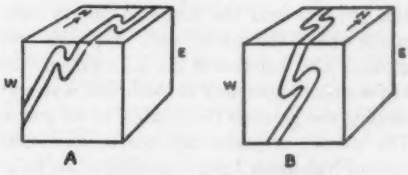


FIGURE 5.—BLOCK DIAGRAM OF FOLDS
A indicates shearing of west side of block upward.
B indicates shearing of east side of block upward.

The thickness of the highly folded and fairly structureless upper member of the Littleton formation is more difficult to determine. The upper member is probably thicker than the lower and middle members, so that 6000 feet may be a fair estimate (Structure section AA′, Pl. 1).

These figures give a grand total of about 17,000 feet of Littleton formation in the Monadnock area. This figure must, of course, be regarded as an approximation.

Igneous Bodies

Oliverian magma series.—The evidence that the granodiorite of the Oliverian magma series is a large concordant body forming the core of the Swanzey dome has been discussed earlier in this paper.

New Hampshire magma series.—Large intrusive bodies of the New Hampshire magma series occur as discordant and concordant plutons throughout the Monadnock area.

KINSMAN QUARTZ MONZONITE: The large discordant body of Kinsman quartz monzonite in the northeastern part of the Monadnock quadrangle is 3½ miles wide and 6½ miles long. The southern boundary is poorly exposed due to lack of outcrops, except in the Bonds Corner region. Here, the general trend of the contact zone is northwesterly, cross-cutting the northerly striking schists of the Littleton formation. North of Skatutakee Lake, 1 mile east of Harrisville, the foliation of the Kinsman quartz mon-

zonite near the contact strikes NE. and dips 40° NW. under the gneiss. Great slabs of quartz-mica schist and gneiss are surrounded by the quartz monzonite, which has a fine-grained granulated nonporphyritic chilled border at this point. The large body of Littleton formation making up Blood Hill and Cobb Hill projects into the Kinsman quartz monzonite which surrounds and cross-cuts the gneiss. The foliation of the quartz monzonite at the eastern boundary of Cobb Hill is steeply dipping and parallels the foliation of the gneiss. The contact of gneiss and quartz monzonite west of Nubanusit Lake is parallel to the foliation of the gneiss with northerly strikes and vertical to steep westerly dips.

The foliation throughout the Kinsman quartz monzonite strikes N. to NNE., with vertical to steep westerly dips. In many places the phenocrysts are aligned parallel to the foliation with the longest axis in a steep to vertical position. Many of these phenocrysts are augen; granulated quartz and feldspar and foliated biotite surround the lenticular crystals.

Roof pendants, a few large enough to show on the geologic map (Pl. 1), are numerous in the Kinsman quartz monzonite. They lie parallel to the regional schistosity with north-south to northeast strikes and steep dips, and do not appear to have moved from their original position. These are labeled *Dl* on the geologic map, although they may belong to the upper member of the Littleton formation. Many smaller inclusions of schist from a few inches to several feet in length are present, generally aligned parallel to the foliation plane and showing all stages of assimilation. Some of the biotitic schlieren may represent partially digested schist.

There are many smaller bodies of Kinsman quartz monzonite southwest of the main intrusive. Some of these have small phenocrysts; others lack them. The largest of these intrusives have been mapped with dotted boundaries where it has been impossible to trace their contacts (Pl. 1). Many are injected parallel to the foliation of the schists of the Littleton formation and have a distinct foliation of their own parallel to the structure of the adjacent schists. Podlike stringers of quartz monzonite are numerous in the schists of the Littleton formation; many of these represent material injected from the Kinsman quartz monzonite.

The structural features of the Kinsman quartz monzonite suggest the intrusion of a body of batholithic proportions in the northeastern part of the Monadnock quadrangle. The chilled border of quartz monzonite dipping under the schists of the Littleton formation and cross-cutting the general strike of the schists, as well as the presence of roof pendants of schist in the higher portions of the intrusive, suggest this theory. There is a chilled-border type of Kinsman quartz monzonite around the large roof pendant in the northeast corner of the map where the schists are apparently in their original position. An upward-moving mass of magma is indicated by the presence of phenocrysts whose longest axes are steep to vertical. The alignment of the phenocrysts parallel to the regional foliation and parallel to the foliation of the quartz monzonite is essentially primary. Later movements are indicated, however, by the development in some places of augen with highly granulated borders. Late orogenic movement caused this extreme granulation and the concentration of biotite into thin layers; these features are more pronounced than one would expect from the flow of a magma.

The coarseness of the Kinsman quartz monzonite suggests slow cooling. Complete absorption of schist inclusions is indicated by an excessive amount of garnet and some cordierite, giving the quartz monzonite a contaminated aspect in many areas. The introduction of pod-like stringers and porphyroblasts of potash feldspar into the surrounding schists, as well as the intense metamorphism of the Littleton formation in this region, may be explained by the closeness of the Kinsman quartz monzonite underlying the schists. The many smaller bodies southwest of the main intrusive represent off-shoots from underlying magma injected as sill-like masses into the overlying Littleton formation.

SPAULDING QUARTZ DIORITE: The Spaulding quartz diorite occurs in bodies of various sizes; some are sill-like, while others have cross-cutting contacts. The large body 4 miles long and 2 miles wide in Marlboro township has a weak foliation that strikes northeast and dips steeply. It forms the core of an anticline.

The contacts are poorly exposed, but it is surmised that the northwest and southeast contacts are concordant; the northeast contact is probably concordant, although later deformation makes it appear to cross-cut the schists of the Littleton formation locally. The southwest contact is so poorly exposed that it is impossible to say whether it is concordant or discordant.

The Spaulding quartz diorite in the vicinity of Gap Mountain is granulated and nonfoliated with irregular contacts cross-cutting the regional structures. Inclusions of schist oriented parallel to the regional foliation, as well as roof pendants in their original position, suggest that the Spaulding quartz diorite exposed here is near the top of the body. This fact is corroborated by the contaminated aspect of the rock, which has many angular and partially digested schist fragments, as well as pseudosillimanite crystals. A high percentage of biotite in knots around granulated quartz-feldspar aggregates also suggests contamination.

Many elongate bodies of Spaulding quartz diorite are found south and east of Mt. Monadnock, as well as in Fitzwilliam, notably east and west of Laurel Lake. These bodies tend to be sill-like and are generally well foliated parallel to the regional foliation, with vertical to steep westerly dips. It has been possible to map only the largest of these, and in many instances lack of outcrops made it impossible to trace their boundaries accurately. They indicate a forceful intrusion into already folded schists, probably before the completion of folding, their foliation being in part primary and in part due to regional compression. The presence of these foliated bodies together with nonfoliated bodies of Spaulding quartz diorite suggests that the Spaulding quartz diorite was intruded over a long time at slightly different periods, the later bodies lacking the strong foliation of earlier bodies. The more foliated bodies resemble the Bethlehem gneiss of areas to the north and possibly should be correlated with it (Hadley, 1942, p. 144). The Bethlehem gneiss is believed to have preceded the intrusion of the Kinsman quartz monzonite. But since the large bodies of Spaulding quartz diorite are less foliated than the Kinsman quartz monzonite in this area, they were presumably intruded after the Kinsman quartz monzonite. Cross-cutting contacts of the two bodies have not been found.

Concord Granite: The Concord granite occurs as a series of large bodies which cut across the existing structures at all angles. Innumerable smaller bodies are present, many dikelike; some are sill-like and were injected parallel to foliation planes. Contacts are generally steep where exposed. The Concord granite is the least foliated of the New Hampshire magma series. Primary foliation can be found by careful study of some of the quarry exposures; secondary foliation, rare, is suggested in some places by extreme granulation and a gneissoid appearance of the granite. Orogeny appears to have been completed, or practically completed, when the Concord granite was intruded. Remnants of the earlier intrusives as well as the schists are found surrounded and cross-cut by the Concord granite. Inclusions are found in all states of assimilation. The Concord granite has various phases which came in at slightly different times, but these are so intimately mixed that there are generally no clear-cut contacts.

Faulting—Silicified Zones

Four silicified zones, slightly en echelon to each other, are exposed in a belt extending from half a mile west of Sip Pond, to 1 mile northwest of Pearly Pond. The rock is almost pure quartz, slightly iron stained and brecciated in places. It breaks into angular blocks which can be followed for a distance of half a mile or more where outcrops are rare. The northernmost of the silicified zones strikes NE. and dips 65° NW., and the more southerly three strike NE. to N.–S., with vertical or steep westerly dips. The zone averages 150 feet wide. The Concord granite and schists of the Littleton formation show intense shearing and granulation for several yards away from the contact on both sides. The trend of the silicified zones is parallel to the regional strike. These silicified zones are presumed to represent faults since they are similar to other silicified fault zones in New Hampshire (Billings, 1941, p. 914).

Age of Structural Features

The main deformation of the Monadnock area is considered to have occurred during middle or late Devonian time. The Littleton formation of lower Devonian age, the youngest sedimentary series in the area, is highly folded and metamorphosed throughout the region. The folding in New Hampshire is pre-White Mountain magma series, which is considered to be Mississippian in age (Billings, 1937; 1941; Kruger, 1946).

The intrusion of the Oliverian magma series preceded the folding of the area, but occurred later than the lower Devonian sedimentation (Chapman, 1939, p. 146). It is younger than the Ammonoosuc volcanics. Angular blocks of the overlying fine-grained, white, fragmental Ammonoosuc volcanics were found in the granodiorite of the Oliverian magma series 0.45 mile southwest of Mt. Huggins at an altitude of 780 feet in the cliffs just south of the trail. Inclusions of the dark-colored Ammonoosuc volcanics in the granodiorite were found 0.8 mile west of Corey Pond. There is no evidence in this area that the granodiorite intrudes the Partridge formation, the Clough quartzite, or the Littleton formation, but Chapman (1939, p. 178) says that the Clough quartzite has been feldspathized by the Oliverian magma series, and therefore the Oliverian magma series is younger than the Clough quartzite. The Oliverian magma series, however, is considered to have been intruded before the folding and metamorphism; the large concordant sheets exposed in the Keene quadrangle would be impossible if the rocks had been folded first. Moreover, the Oliverian magma series is cut by amphibolite dikes, showing that it is pre-metamorphism.

The intrusion of the rocks of the New Hampshire magma series occurred toward the close of the main period of deformation when the rocks of the Littleton formation had already been folded. Granulation in the Kinsman quartz monzonite and in some of the bodies of Spaulding quartz diorite indicates that deformation was still going on. The Kinsman quartz monzonite is considered older than the Spaulding quartz diorite because of its characteristic granulation, the augen character and alignment of the phenocrysts, and the foliation shown by the biotite. Many of the Spaulding quartz diorite bodies appear less granulated and less foliated. Deformation had almost ceased when the Concord granite was intruded, for it is the least deformed of the intrusives; it cuts the other bodies. Most of the pegmatites are undeformed and are therefore considered to be associated with the New Hampshire magma series.

Later faulting in the area is indicated by the presence of four discontinuous belts of a silicified zone which sheared the schists and granites in their immediate vicinity. There is no way of accurately dating this faulting, but from its relation with other areas, it may be considered to be late Devonian, Mississippian, or even Triassic, since it cuts the youngest rocks of the New Hampshire magma series. It is presumably the same age as the Pine Mountain fault of the Mt. Washington area (Billings, 1941, p. 919).

Summary of Geologic History

The oldest rocks of the Monadnock area are the Ammonoosuc volcanics of probable late Ordovician age. Over 600 feet of pyroclastics accumulated in the area. Shale representing the Partridge formation was laid down on top of the Ammonoosuc volcanics in upper Ordovician time, but erosion following deposition has removed most of it; absent in most places, a mere 50 feet thickness is found in a few localities.

Following uplift and erosion, Silurian seas invaded the area; conglomerate and sandstone that eventually became the Clough quartzite accumulated to a depth of several hundred feet. Uplift and erosion following deposition, as well as unequal sedimentation, may account for the thinness of the Clough quartzite in the region.

During early Devonian time, over 17,000 feet of sandstone and shale accumulated in shallow seas which invaded the area. These sediments became the Littleton formation.

Following the deposition of the Littleton formation, the whole area underwent intense deformation in middle or late Devonian time. The intrusion of the granodiorite of the Oliverian magma series into the Ammonoosuc volcanics in the western part of the area probably preceded the deformation. The shale

and sandstones of the area were thrown into broad anticlines and synclines by the orogeny, some areas becoming more tightly folded than others. The schistosity and axial planes of the region took on northeast strikes and steep dips; the axes of the folds plunged north; and the northwest limbs of the synclines were overturned toward the southeast.

During the later stages of the folding, basaltic dikes and sills, later metamorphosed to amphibolite, were intruded into the Oliverian magma series as the earliest phase of the New Hampshire magma series. Then came the intrusion of large bodies of the New Hampshire magma series. First of these, the Kinsman quartz monzonite, has some primary foliation, but a good deal of its gneissoid texture, granulation, and augen structure were developed by the forces of deformation which had not ceased at the time of its intrusion. The presence of folded, foliated inclusions of the Littleton formation in all of the rocks of the New Hampshire magma series is proof that the sediments had already been folded and metamorphosed. The Spaulding quartz diorite closely followed the intrusion of the Kinsman quartz monzonite, showing by its regional foliation that deformation had not ceased when it invaded the region. The Concord granite, the last of the New Hampshire magma series, was intruded when orogeny had almost ceased; it is the least deformed of the New Hampshire magma series.

Metamorphism of the sediments of the area accompanied the deformation and igneous intrusions. The metamorphism is high grade throughout the region; recrystallization has progressed from the beginning of deformation and continued throughout the folding and during the intrusion of the New Hampshire magma series, when heat and gaseous materials were available. Some minerals such as sillimanite were entirely changed to muscovite during the later stages of recrystallization. Chloritoid developed as a retrograde mineral.

Later faulting of the area occurred in late Devonian, Mississippian, or Triassic time, as indicated by four belts of a silicified zone.

Post-Carboniferous uplift and erosion with peneplanation in Tertiary time, followed by more uplift and stream entrenchment before the area was covered by the Wisconsin Ice Sheet, brings the history of the region up to the present. Erosion is going on today.

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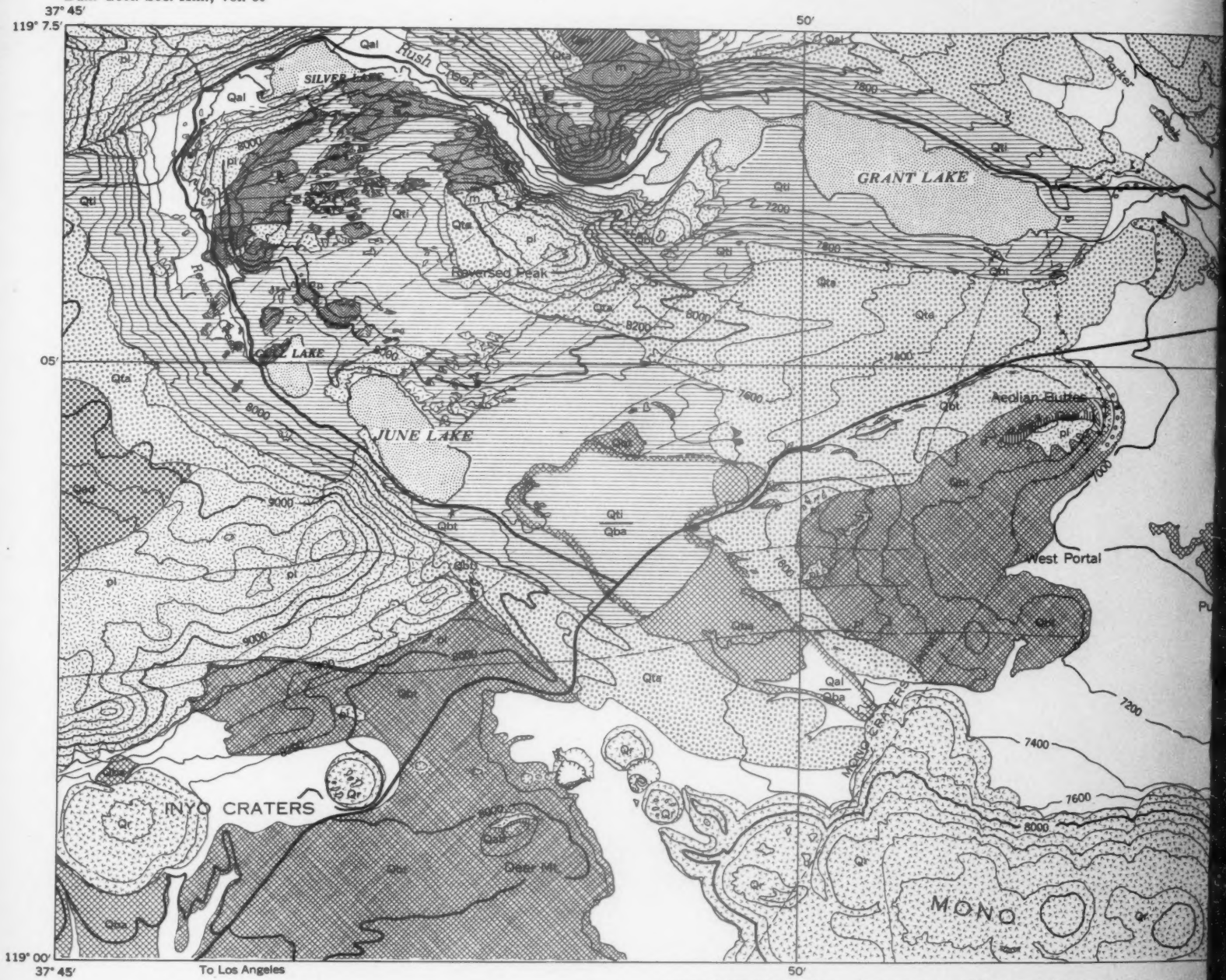

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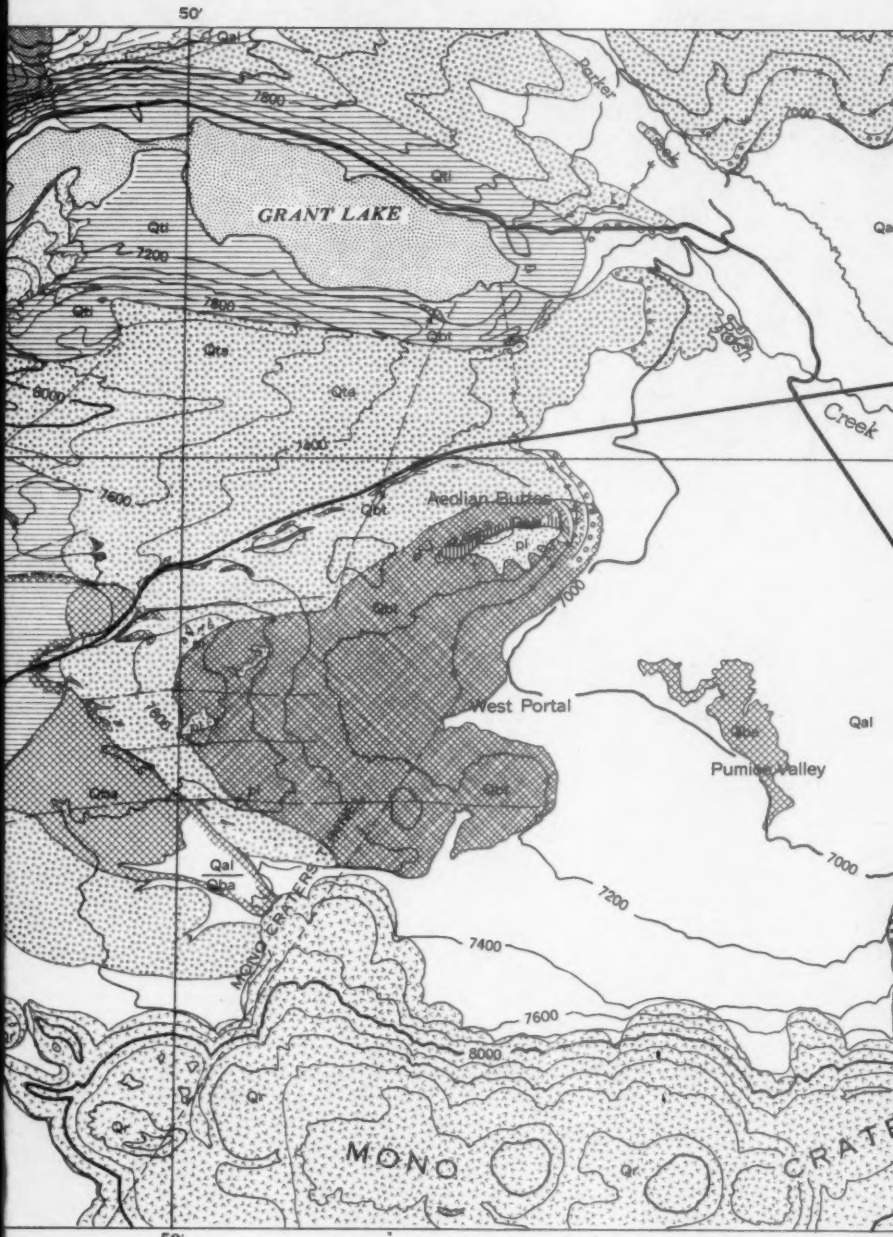

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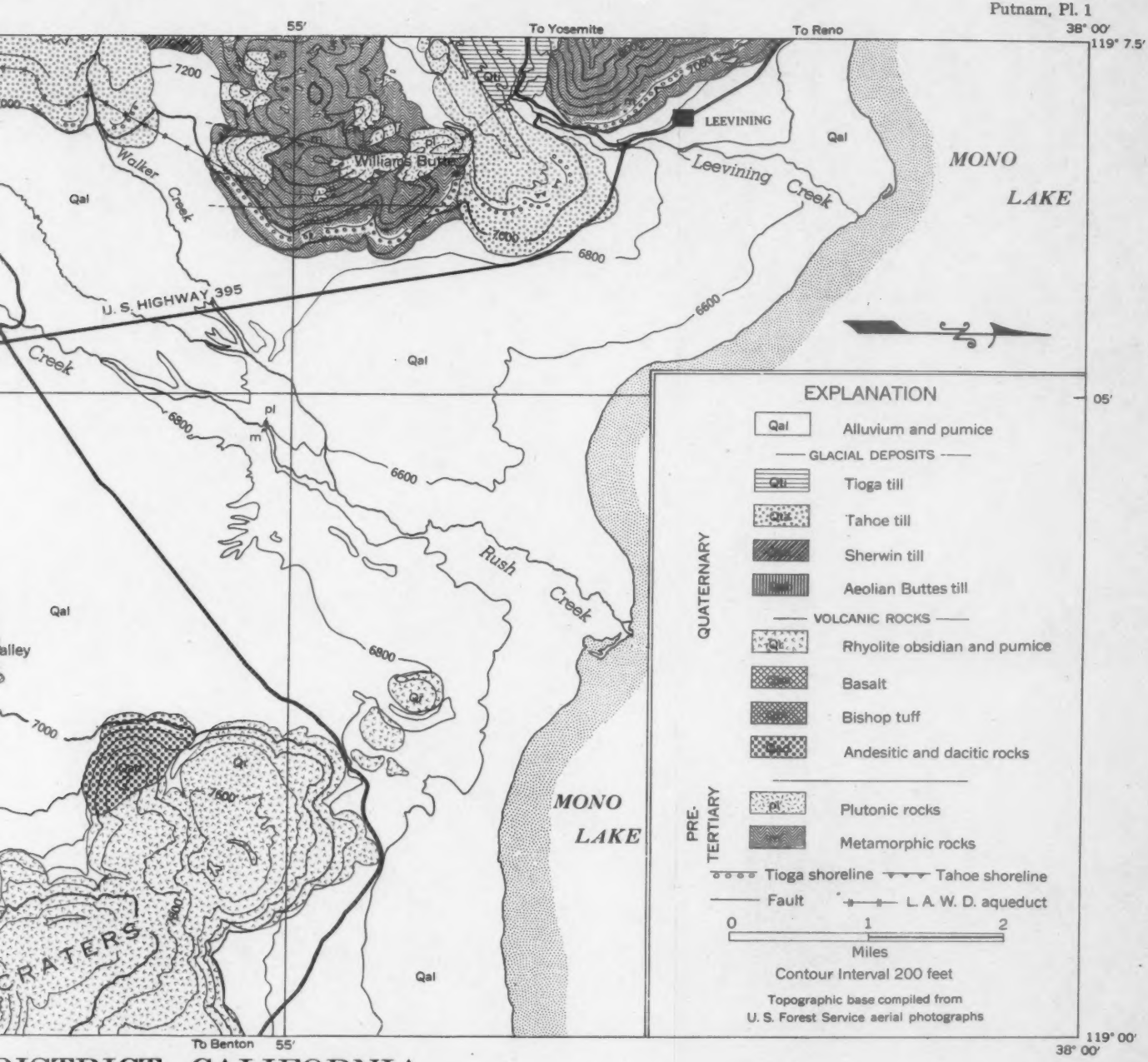

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QUATERNARY GEOLOGY OF THE JUNE LAKE DISTRICT, CALIFORNIA

BY WILLIAM C. PUTNAM

CONTENTS

	Page
Abstract	1281
Introduction	1282
General statement	1282
Regional setting	1282
General description	1282
The Sierra Nevada	1282
The Morainal belt	1283
The Lake Plains	1283
Mono-Inyo craters	1283
Climate	1283
Vegetation	1284
Previous work	1285
Present investigation	1285
Acknowledgments	1286
Rock units	1286
General statement	1286
Pre-Tertiary series	1286
Metamorphic rocks	1286
Intrusive rocks	1287
Quaternary series	1288
Andesitic and dacitic rocks	1288
Aeolian Buttes till	1289
Bishop tuff	1289
Sherwin till	1290
Tahoe till	1290
Basalt	1291
Tioga till	1291
Mono-Inyo craters obsidian	1292
Pumice and alluvium	1293
Structure	1294
Geomorphology	1295
Lake Russell	1295
Mono Lake	1295
Shore lines	1295
Lake sediments	1296
Origin of Reversed-Rush Creek Canyon	1297
Engineering Geology	1299
Mono craters tunnel	1299
Summary and conclusions	1301
References cited	1302

ILLUSTRATIONS

Figure	Page
1. Location of June Lake district	1282
2. Topographic sub-provinces	1282
3. Vertical distribution of life zones	1285
4. Rock units of the June Lake district	1286
5. Rush Creek glacier during the Tahoe stage	1298
6. Geologic cross-section of the Mono Craters tunnel	1300

Plate	Facing Page
1. Geologic map of June Lake district	1281
2. Geographic map of June Lake district.	1286
3. Geomorphic map of June Lake district	1290
4. Aeolian Buttes	1292
5. Volcanic rocks	1293
6. Reversed Creek-Rush Creek trough	1296
7. Glacial and lacustrine deposits	1297

Table	
1. Climate data	1284

ABSTRACT

In the June Lake district of east-central California, the Sierra Nevada has been uplifted along a number of closely-spaced faults, and the rectilinear outline of the range front is controlled by their pattern. That repeated movement is characteristic of these faults is shown by increasing throw where they cut successively older moraines.

There were at least four ice advances in the Pleistocene: the earliest stage, here named the Aeolian Buttes, was followed by the Sherwin, Tahoe, and Tioga, previously named by Blackwelder. Constructional forms have been destroyed on the two older tills but are preserved on the later ones.

Large quantities of rhyolitic ash and pumice, now the welded Bishop tuff, were erupted in the interval between the Aeolian Buttes and Sherwin glacial stages. A small basalt flow near June Lake rests on Tahoe till and was glaciated in part by Tioga ice. Plug domes and flows of rhyolitic obsidian of the Mono and Inyo craters were erupted following the Tioga stage. Pumice from these volcanoes covers late recessional moraines, and these volcanic slopes near Mono Lake are not cut by Pleistocene shore lines.

Mono Lake was much larger in the Pleistocene, and the name Lake Russell is proposed for this lake. During the Tahoe stage the lake had an outlet to the

southeast and was connected with the Owens Valley-Death Valley system of lakes and channels.

Introduction

General Statement

June Lake district in east-central California provides information on the glacial chronology

FIGURE 1.—LOCATION OF JUNE LAKE DISTRICT

of the Sierra Nevada and the relation of this glacial history to fluctuations of Mono Lake, a large permanent lake of the Basin and Range Province. The region is noteworthy for its abundant Recent and Pleistocene volcanism, with the unusual feature that flows and tuffs are interbedded with glacial till. This alternation of volcanic rocks with glacial deposits makes certain the existence of four glacial stages in the area.

Regional Setting

General description.—The June Lake district, about 350 miles north of Los Angeles and 180 miles east of San Francisco, includes the eastern slope of the Sierra Nevada and the western margin of the Basin and Range Province. This is a region of strong relief, with bold precipitous slopes in the Sierra and desola sage plains and a near-lifeless carbonate la in the desert section. The district can b divided into four topographic subprovinces (F 2): (1) the Sierra Nevada, (2) the morain

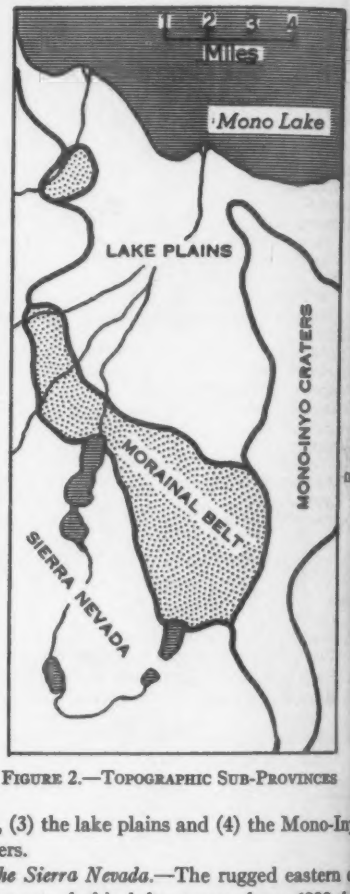

FIGURE 2.—TOPOGRAPHIC SUB-PROVINCES

belt, (3) the lake plains and (4) the Mono-Inyo craters.

The Sierra Nevada.—The rugged eastern escarpment of this lofty range, here 6000 feet higher than the plain at its base, forms the western margin of the area. Slopes are steep to nearly vertical and there is much bare rock. All the larger valleys have been glaciated; tarns and morainal lakes are numerous. Divides are narrow, as a rule, and rock pinnacles and craggy summits are common, especially on granitic and metamorphic rocks. The principal exceptions are ridges of volcanic rock or those underlain by older glacial till; both of

these are generally rounded. Summit uplands are moderately well-developed in the Upper Alpine Zone, and most are blanketed with frost-shattered rock. The abrupt break in slope between these summit uplands and the precipitous canyon walls is a striking feature of the range.

The profile of the eastern face of the Sierra Nevada does not slope continuously from base to crest but is more nearly comparable to a monumental stairway with bench-like shoulders separated by steep slopes. In plan, the escarpment is not a straight line but makes a rectilinear pattern of blunt salients and deep re-entrants. There are no foothills in the ordinary sense of the word; the transition from plain to mountain is abrupt.

The Morainal belt.—Mouths of larger canyons are flanked by lateral moraines that make a series of crescentic ridges extending beyond the base of the Sierra. Moraine crests are narrow, slopes are broad, and these embankments rise to altitudes as much as 800 feet above the lake plains they enclose. Lateral moraines are many times larger than the relatively diminutive terminal moraines. In fact, terminal moraines of the next-to-last glacial stage (Tahoe) no longer survive, and terminal moraines of the last glacial advance (Tioga) are nested inside bulkier lateral moraines of the preceding (Tahoe) stage.

The morainal belt makes a striking contrast with the more rugged Sierra Nevada. Slopes are gentler and their mantle of gray-green sage contrasts strongly with the somber green of the pine forest or variegated red, brown, and gray outcrops of the higher mountains. The great volume of these moraines is compelling evidence of the effectiveness of glacial erosion; they represent the removal and deposition of many thousands of cubic yards of rock.

The Lake Plains.—The broad sage plain laid bare by the receding waters of glacial Mono Lake is in the central part of the area. This plain is not level but terraced, and these terraces are conspicuous features to an altitude of about 650 feet above the present lake. They are cut on moraines of the next-to-last (Tahoe) glaciation, as well as the rock slopes of the Sierra Nevada. The town of Leevining stands on a wide lake terrace 380 feet above the lake, and this same post-glacial terrace makes the broad and desolate expanse of Pumice Valley.

Sierran streams, such as Rush and Leevining creeks, that empty into Mono Lake have cut 300-foot canyons through lacustrine silt, sand, and gravel with the lowering of lake level beginning with the retreat of the Tioga glaciers.

Mono-Inyo craters.—These rhyolitic obsidian domes and flows, one of the more striking features of the region, make a nearly continuous chain from Mono Lake southward along the eastern border for 19 miles. The individual domes are nearly circular in plan, with pinnacled summits and steep slopes screened by talus aprons. The flows are stubby, with precipitous sides sometimes 300 feet above the plain, and the surfaces are craggy and rough. Like the domes, they are flanked by a virtually impassable talus blanket of sharp-edged obsidian blocks.

Explosion pits are common, and vast quantities of pumice and obsidian fragments have been hurled from these vents. Pumice extends at least as far as the Nevada boundary, 30 miles east. To the west, or to windward, pumice is not as widespread but it obscures most of the central part of the area and covers late recessional moraines of the last glaciation.

Climate

The climate is continental—cold and snowy in winter, dry in summer with warm days and cold nights. The summer drought is broken by frequent thunder showers, especially at higher altitudes. Precipitation decreases markedly with distance from the mountains and lower altitudes.

The characteristics of the climate, lower temperature and increased precipitation with higher altitude, are shown by Table 1. Cain Ranch is near the junction of U.S. Highway 395 and the June-Grant Lake road. Gem Lake is on Rush Creek immediately west of the area mapped, and Ellery Lake is at the head of Leevining Canyon on the Tioga Pass road to Yosemite Valley.

The climate follows the general pattern of central California with the greatest precipitation in winter and near-drought in summer. In summer, the strong difference in temperature

between the desert plains bordering Mono Lake and the summit region of the Sierra Nevada is especially striking.

The significant geological effects of the climate are the great effectiveness of frost work at higher altitudes and the relatively slight role of chemical weathering except on the more vulnerable rocks, such as lime-silicate hornfels. Soil-forming processes are much less effective than on the western slope of the Sierra Nevada and fresh rock in outcrops is the rule rather than the exception. There are wide areas of glacially-polished rock.

TABLE 1.—CLIMATIC DATA

	Cain Ranch* Elev. 6854		Ellery Lake† Elev. 9600		Gem Lake** Elev. 9030	
	Av. Temp.	Av. Precip.	Av. Temp.	Av. Precip.	Av. Temp.	Av. Precip.
January	24.4	1.97	22.9	5.04	26.1	4.60
February	26.0	2.63	23.2	5.85	27.1	5.33
March	33.7	1.44	27.7	4.02	32.5	3.21
April	42.1	0.48	32.0	2.88	38.3	2.17
May	49.4	0.37	38.6	1.35	45.9	0.91
June	56.5	0.44	46.9	1.29	53.1	0.71
July	64.0	0.26	55.0	0.69	60.3	0.57
August	62.8	0.40	54.6	0.70	59.9	0.52
September	55.1	0.22	47.5	0.54	52.3	0.49
October	45.5	0.75	39.5	1.37	44.4	1.42
November	35.1	0.70	32.0	2.14	36.1	1.44
December	28.5	1.98	26.4	4.96	29.7	4.20
Annual	43.6	11.64	37.7	30.83	42.1	24.57

* 14 years record
† 21 years record
** 21 years record
Data supplied by Weather Bureau, U.S. Department of Commerce, Los Angeles Office, letter of October 30, 1946.

Vegetation

The vegetation pattern is especially striking because of the abrupt changes of climate within short distances, both laterally and vertically. All life zones from Upper Sonoran to Artic-Alpine are present but are much compressed because of the several narrow climatic zones on the steep, east-facing Sierran escarpment. Several factors control this vegetation pattern, among them: (1) vertical zonation due chiefly to temperature and effective precipitation, (2) cold-air drainage down canyons causing the intrusion of higher altitude forms into a lower thermal zone, (3) the nature of the soil, chiefly as its affects the available moisture, and (4) direction of slope, whether dominantly shady or sunny.

The control of altitude is shown graphically on the idealized transect (Fig. 3) from Cain Ranch to Mt. Lewis. These data were kindly given me by Mr. Elden Vestal, Fisheries Biologist, California Division of Fish and Game. Principal indicators for the life zones are:

Life Zone	*Indicator*	*Life Form*
Arctic-Alpine	*Dodecatheon alpinum*	Herb
	Erigeron salsuginosus	Perennial herb
Hudsonian	*Pinus balfouriana*	Forest tree
	Pinus flexilis	
Canadian	*Pinus murrayana*	
	Abies concolor	
	Populus tremuloides	
Transition	*Pinus jeffreyi*	
	Purshia tridentata	
Upper Sonoran	*Artemisia tridentata*	Shrub
	Pinus monophylla	Small tree

The factor of air drainage complicates this essentially simple picture. For example, Lodgepole pine (*Pinus murrayana*) grows as low as 7000 feet in canyons where cold air drainage is confined. One other notable zone-crosser is aspen (*Populus tremuloides*) which grows in spring zones and along stream banks through a vertical range from 6400 to 10000 feet.

The control of the soil on vegetation is strikingly evident in aerial photographs; a characteristic vegetational assemblage within a life zone can commonly be related to a given rock type. For example, in the Transition Zone, lateral moraines support only a sparse growth of Jeffrey pine and are characteristically mantled by sage (*Artemisia*), which gives them a desolate and highly distinctive appearance. Surprisingly enough, in the same life zone, pumice from the Mono and Inyo craters supports one of the more flourishing groves of Jeffrey pines in the region. These large trees suffer from a fatal handicap in this terrain. Pumice gives inadequate anchorage and when these trees reach maturity they are readily toppled by the wind. "Fallen forests" are common and

tree trunks generally show a preferred orientation in the direction of the prevailing wind. Windfalls make a distinctive pattern on aerial photographs and at first glance can be confused with lineations in the bedrock.

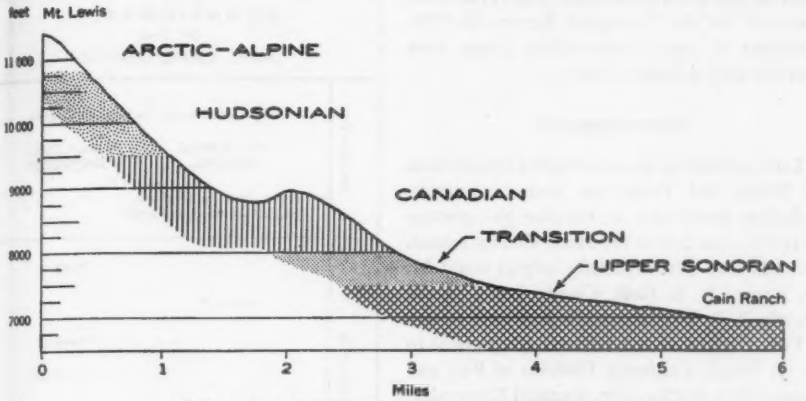

FIGURE 3.—VERTICAL DISTRIBUTION OF LIFE ZONES

In places granitic rocks can be distinguished from metamorphic rocks by their more open "speckled" forest of pines; the less broadly fractured metamorphic rocks support a denser flora of juniper, shrubs, and grass. In fact, the fracture pattern locally controls an alignment of trees along fault lines in granite, probably as a result of the damming of ground water along these planes.

Previous Work

This region attracted the attention of geologists from the pioneer period, and a number of competent observers have visited the area. Perhaps the first white men to visit the region were Joseph R. Walker and his party who came through in 1833 (Farquhar, 1925, p. 8). There was a flurry of geologic activity shortly after California's admission to the Union. The Whitney Survey visited the area, made a number of observations on the Mono Craters, and explored the Sierra Nevada in an attempt to locate a pass across that formidable barrier (Brewer, 1930, p. 416). Both J. D. Whitney (1865, p. 454–455) and Joseph Le Conte visited the Mono Craters and the latter correctly interpreted the rounded cobbles he found on their summits as accidental ejecta (Le Conte, 1879).

The outstanding achievement in the geologic exploration of the eastern Sierra Nevada and the Mono Basin was the work of I. C. Russell (1889) and his topographer, W. D. Johnson. Russell's scientific contributions were outstanding; they include an accurate description of the landforms of the region, a recognition of multiple glaciation, evidence for more than one expansion of Mono Lake, for the importance of faulting in the elevation of the Sierra Nevada, and a quite detailed analysis of the mechanism of ice erosion and of the type of volcanic eruption represented by the Mono Craters.

Because of this achievement, I propose that Pleistocene Mono Lake be named Lake Russell, and this usage is followed in this paper. I believe this is preferable to Russell's suggestion of using Mono Lake for the existing lake and Lake Mono for the Pleistocene water body.

Present investigation

Field work started in the summers of 1936 and 1938, interrupted by World War II, was completed in the summer of 1946. Mapping was done on U.S. Forest Service aerial photographs using a military-model pocket stereoscope in the field. The topographic base for the geologic map (Pl. 1) was made by using a a mirror stereoscope to locate contours on the photographs and a vertical sketchmaster to

transfer them to a U.S. Forest Service planimetric map, scale 1:31,680. Vertical control for the low country was established from a survey made by the Los Angeles Department of Water and Power. Altitudes of higher peaks were taken from the Mt. Lyell quadrangle (scale, 1:125,000) surveyed by the Geological Survey in 1898. Altitudes of many intermediate points were checked with a Paulin altimeter.

Acknowledgments

I am indebted to the Los Angeles Department of Water and Power for many courtesies, including permission to examine the geology in the western half of the Mono Craters tunnel. Officials that were especially helpful were: M. D. Burris, L. E. Goit, Claude James, S. B. Morris, R. R. Proctor, and H. A. Van Norman.

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Rock Units

General Statement.

The rocks are separable into two categories: (1) a bedrock series of metamorphic and igneous rocks overlain by (2) Quaternary glacial till, stream and lake deposits, and volcanic flows, tuff, and pumice. Glacial deposits alternate with volcanic material, so that most of the volcanic rocks except for the andesitic and dacitic types, can be dated in terms of glacial stages, and, therefore, the Quaternary rocks are described in chronologic sequence.

Pre-Tertiary series

Metamorphic rocks.—The metamorphic rocks are hornfels and crop out on the bold ridge immediately west of Leevining, on Williams Butte, on the west flank of Reversed Peak, and are exposed in the Mono Craters tunnel east of the Mono Craters.

These are meta-sediments of great variety. Calc-silicate hornfels are white or bluish-gray rocks that make conspicuous outcrops. Chert layers are characteristic and are more resistant

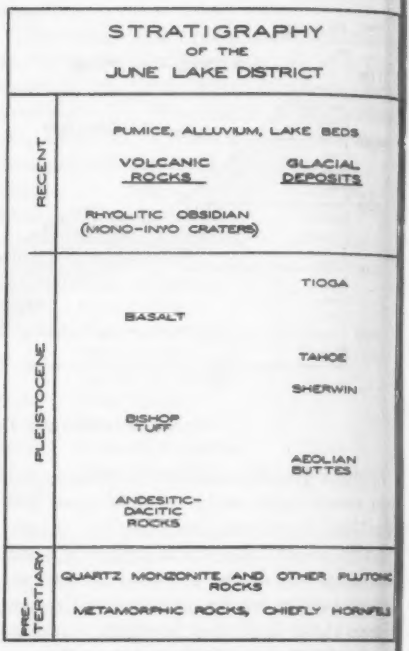
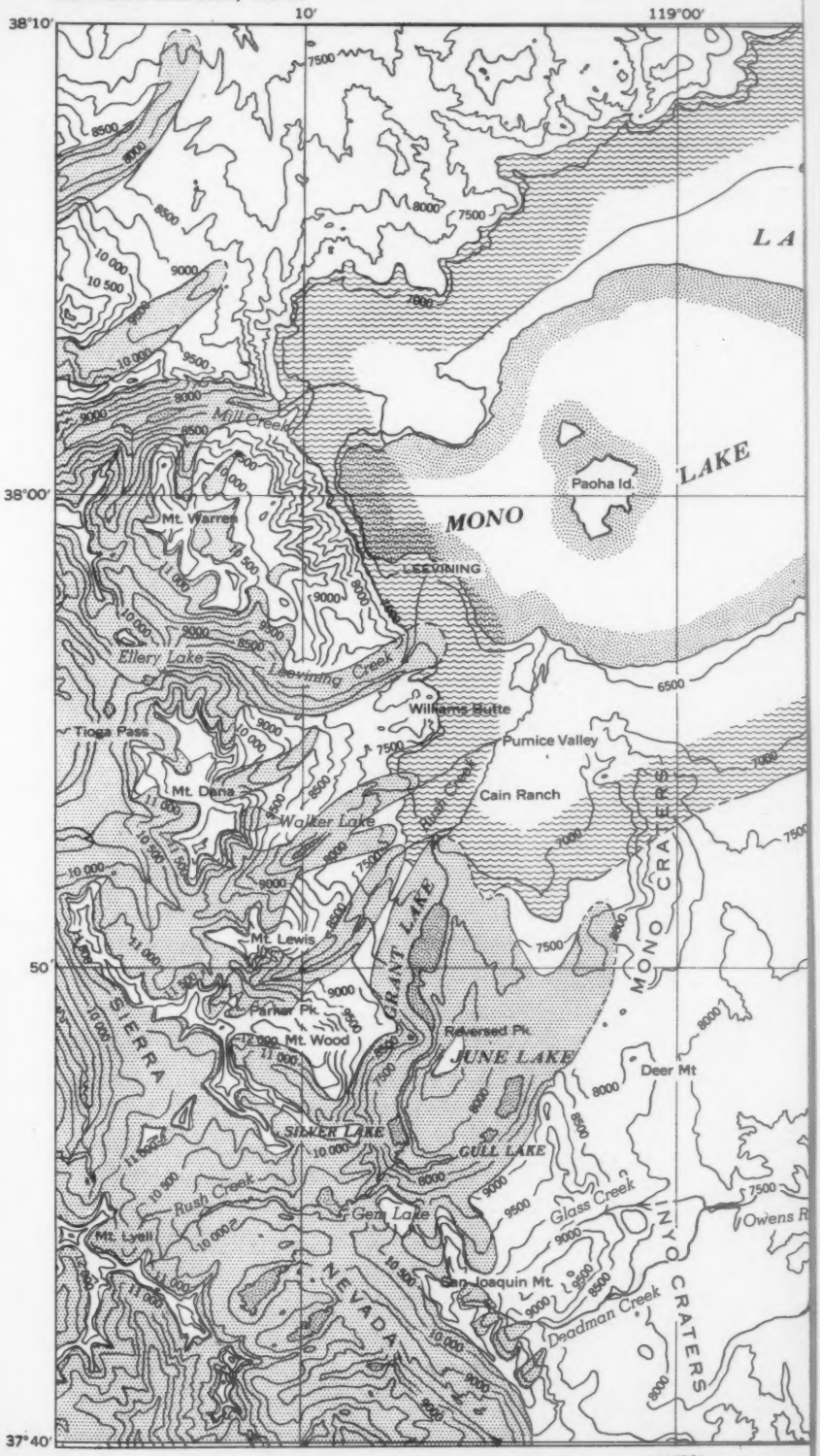


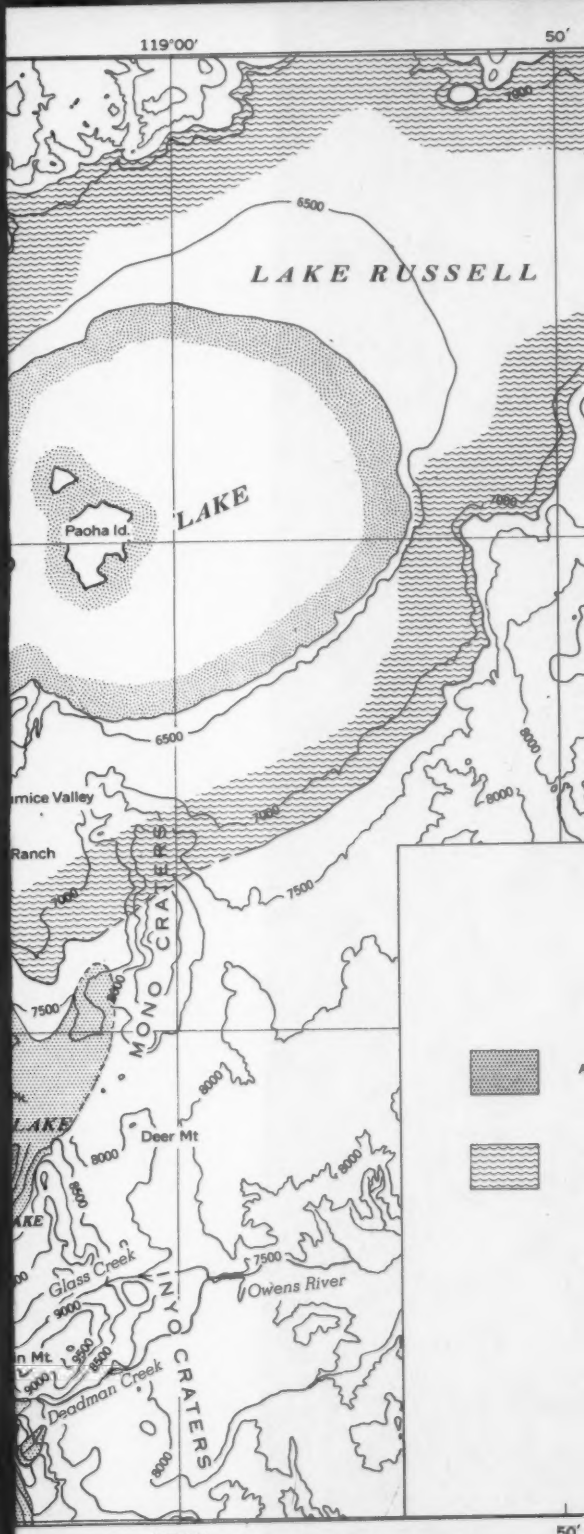
FIGURE 4.—ROCK UNITS OF THE JUNE LAKE DISTRICT

to weathering; they stand out and give the rock a gnarled surface. Some varieties appear to be quartzite, but actually are recrystallized limestone and probably contain tremolite and other calcium-magnesium silicates. Garnetiferous rocks are found near the contact with quartz monzonite on Reversed Peak, and elsewhere in the calc-silicate hornfels. The calc-hornfels is thickest near the instrusive contact at Gull Lake, and crops out in a narrow band on the west side of Rush Creek near Grant Lake.

Associated with the calc-silicate hornfels are resistant, banded, and variegated rocks, probably recrystallized limy shale and sandstone. Some metamorphosed sandstones still preserve relic cross-bedding and what appear to have been detrital grains, especially in more limy



GEOGRAPHIC MAP OF JUNE LAK



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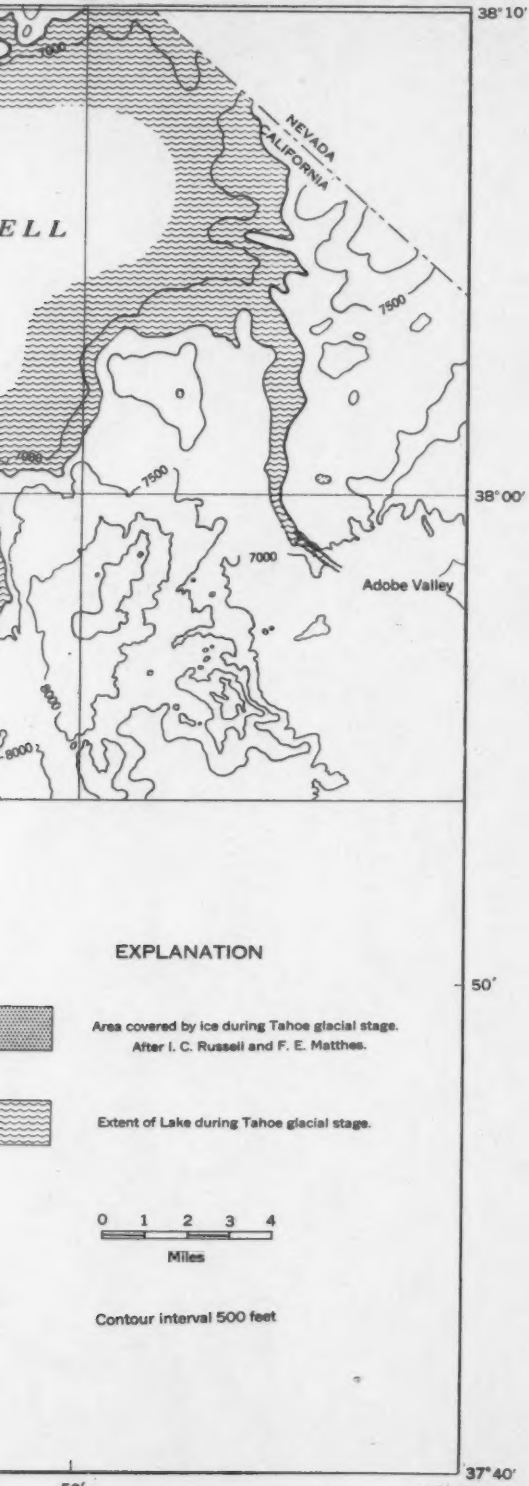


ICT. CALIFORNIA

facies of the sandstone. These metamorphosed sandstones are commonly blue-gray and locally include light-gray to nearly white layers.

Banded hornfels are among the more resistant rocks of the region. Where eroded by ice they make prominent bosses and *roches moutonées*. These are especially well-developed in the valley of Reversed Creek immediately west of Gull Lake and on the west side of Rush Creek narrows.

Shale hornfels is common in Rush Creek valley between Silver Lake and the narrows, on Williams Butte and on the ridge west of Leevining. This rock is dense, closely-jointed and dark: black, red, and dark brown are the usual colors. Sandy facies commonly are lighter-colored; some varieties are purple-brown and quite micaceous, others are silver-gray and grittier.

The main body of metamorphic rocks is cut by many small intrusions of quartz monzonite, quartz diorite, diorite, and gabbro; most are too small to be shown at map scale. The more basic, especially in the Reversed Peak area, appear to be sills; others are cross-cutting, as on Williams Butte where the most common igneous rock is quartz monzonite.

The hornfels has a wide range in erosional resistance; banded varieties are among the more resistant rocks of the district, recrystallized shale among the less resistant. Spacing of joints appears to be one of the controlling factors in resistance to weathering. Shale hornfels behaves much like the parent rock; where the attitude is steep, as it is over most of the area, strike ridges separated by narrow subsequent ravines are common. This belted topography is conspicuous on the west end of Reversed Peak ridge between Silver and Gull lakes. This ridge was strongly glaciated and ice erosion was markedly selective; weak beds were quarried and resistant ones were left to make narrow ridges. The dip here is nearly vertical, the strike almost at right angles to the direction of ice advance. The resulting topography looks as though much of the country had been harrowed.

Calc-silicate hornfels and banded hornfels are nearly as resistant as quartz monzonite. Where glaciated, these rocks make knobs and bosses with domed summits and nearly vertical sides. These large *roches moutonées* are prominent in Reversed Creek valley near Gull Lake.

Hornfels in nonglaciated areas, such as Williams Butte and Leevining ridge, is less resistant than quartz monzonite and makes sharper-crested divides and seemingly narrower canyons. Slopes are mantled with a nearly continuous carpet of small and strongly angular joint blocks. Granitic rocks on Williams Butte make bouldery or rounded outcrops that stand out in bold relief when compared to the surrounding hornfels.

Intrusive Rocks.—Several varieties of intrusive rocks were separated in the field, but are not differentiated on the geologic map (Pl. 1). Quartz monzonite is most abundant, followed by hornblende-biotite-quartz diorite, and smaller bodies of diorite, hornblende gabbro, and gabbro. Quartz monzonite probably makes up about 90 per cent of the exposed plutonic rock. The two largest bodies underlie (1) the outer margin of the range, including outliers such as the Aeolian Buttes and Williams Butte, and (2) the ridge west and south of Silver Lake.

The quartz monzonite is a medium- to coarse-grained rock consisting of orthoclase, plagioclase, quartz, biotite, and hornblende. Orthoclase is euhedral, pinkish, and in places is in large crystals, giving the rock a markedly porphyritic texture.

Quartz monzonite commonly has sharp contacts, even against earlier intrusions such as quartz diorite. However, the rock locally contains many xenoliths, including quartz diorite. Perhaps the most noteworthy example of contaminated rock is southwest of Silver Lake where the quartz monzonite includes a large volume of quartz-feldspar porphyry. The dense, dark-gray to black groundmass of this rock contains euhedral feldspar phenocrysts up to 0.5 inch across and transparent, anhedral quartz. The volume of included rock, in some places, appears to be equal to, if not greater than, the host. Elsewhere the porphyry is shredded away until only wisp-like stringers survive in the quartz monzonite. Although the origin of the porphyry is uncertain, it may be a meta-volcanic.

Hornblende-biotite-quartz diorite is marginal to the much larger intrusion of quartz monzonite. Perhaps the most extensive outcrop is in

Rush Creek narrows. Soda-lime feldspar is clearly more abundant than potash feldspar; quartz is moderately abundant. Greenish-black hornblende and biotite make a strong contrast to the gray-white background of quartz and feldspar.

Diorite and gabbro make minor concordant intrusions, chiefly in the metamorphic belt on the west shoulder of Reversed Peak. Contacts are ill-defined and these rocks are much contaminated, especially where they are intrusive into calc-hornfels. Basic and intermediate rocks are earlier than the quartz monzonite and quartz diorite and are themselves recrystallized. They are strongly epidotized; their greenish color is perhaps their most distinctive characteristic. Minerals are no longer sharp and well-defined and almost all have blurred margins.

Quartz monzonite underlies the highest peaks in the area, such as Reversed Peak, the 10,800-foot mountain south of Rush Creek and the prominent ridge west of the Inyo Craters. Rock outcrops differ greatly in glaciated and nonglaciated areas (Pl. 6 ,fig. 2).

Where the joint interval is large, quartz monzonite strongly resisted glacial erosion. It was readily quarried where the spacing is close, and it was most susceptible to erosion along faults. Although displacement on individual faults may be slight, and gouge or shattered rock is rare, selective quarrying has etched out some of these features strikingly (Pl. 6, fig. 2). Glacial polish is rare but locally well-developed, and there are moderately extensive areas on the ridge between Gull and June Lakes and on the steep slopes down which Rush Creek cascades southwest of Silver Lake.

Nonglaciated quartz monzonite that lies below the Artic-Alpine Zone weathers into boulders, some of which are left perched on a barren pavement, especially in the Upper Sonoran Zone. The quartz monzonite disintegrates almost directly to sand, which is removed by rill wash nearly as rapidly as its forms so that new surfaces are continually being exposed to the air.

The petrographic description of the Half Dome quartz monzonite (Calkins, 1930, p. 126) reads as though it were a similar rock. But the quartz monzonite of the June Lake district has produced no landforms comparable to the Yosemite domes, presumably because of the close spacing of joints here compared to the wide interval at Half Dome. One of the dominant sets is generally near-vertical, in many places quite close- and uniformly-spaced (Pl. 4, fig. 1), with an interval of 1–3 feet.

Quartz monzonite locally develops a case-hardened weathered surface; elsewhere it is roughened and granular, and in a few places cavernous. Orthoclase, especially in the markedly porphyritic varieties, is resistant relative to other minerals and has moderately strong relief. Inclusions of mafic minerals are quite resistant and project as knobs. In places the joints of deeply-weathered rocks cemented by iron oxide or silica stand out in contrast to the adjacent rock.

Quaternary series

Andesitic and dacitic rocks.—These rocks are exposed in two widely separated areas. One, a biotite dacite, is on the ridge southeast of Reversed Creek; the other, a hornblende andesite, is near the northern end of the Mono Craters. Although the age of neither rock has been determined satisfactorily, both may have been erupted in the early Pleistocene.

The dacite is a porphyritic flow rock, with biotite and quartz phenocrysts in a brown to purplish-gray ground mass. It very likely rests on quartz monzonite, and possibly in part on hornfels. The dacite has been stepped down by faults of the Silver Lake system; to the southwest of the area it is cut off against granite and metamorphic rocks exposed in the escarpment of San Joaquin Mountain. These volcanics are preserved in the summit region of the upthrown side and are separated by a throw of about 400 feet from those on the lower ridge.

The hornblende andesite makes a conspicuous 1000-foot dome near the north end of the Mono Craters and is an anomaly, both petrographically and morphologically, in this chain of obsidian plug domes and flows. Although it is a dome, the sumit lacks the characteristic spines and crags of the obsidian volcanoes and appears to be made of superimposed, moderately viscous flows.

The andesite is black or very dark brown. The dark aphanitic ground mass contains large white

feldspar crystals. The rock has a crude columnar structure and a pronounced tendency to cavernous weathering. Its age is difficult to determine but it is older than the Mono Craters obsidian, since it is partially buried by flows and pumice. The andesite dome is also older than the Tioga expansion of Lake Russell because it is cut by shore lines while the Mono Craters are not. The bedrock contact is hidden beneath pumice, concealing the relationship to the Bishop tuff (mid-Pleistocene) exposed near by.

Aeolian Buttes till.—The oldest recognized till crops out on the crest of the Aeolian Buttes, a low craggy ridge between U. S. Highway 395 and the Mono Craters. Here, in an erosional window cut in the Bishop tuff, the till can be seen resting on a planed and probably glaciated surface of quartz monzonite. The till is also encountered in the Mono Craters tunnel beneath the Bishop tuff. There is little doubt as to its glacial origin here, since poorly sorted angular blocks of a wide variety of rocks, many of them as much as 10 feet across, are embedded in a clay-silt matrix.

The till exposed at the surface consists of scattered, angular rock fragments up to 4 feet across, including a variety of hornfels types and quartz-feldspar porphyry. Quartz monzonite blocks are also present, but the bedrock on which the till rests is an abraded surface of quartz monzonite. None of the other rocks in the till crop out in the immediate vicinity and most of them must have been transported several miles.

That the till underlies the Bishop tuff is almost certain both from evidence in the tunnel and the nearly conclusive relationship on the Aeolian Buttes, where till rests on quartz monzonite and is rimmed by outward-dipping, unglaciated Bishop tuff (Pl. 4, fig. 2).

Deer Mountain, east of Highway 395 and between the Inyo and Mono craters, is the other occurrence of Aeolian Buttes till in the district. Here, angular blocks which include hornfels and granitic rocks rest on quartz monzonite. The age relationship is less certain because the Bishop tuff has been stripped from the hill top. The deposit in this area was beyond reach of ice during the last two glacial stages, and presumably during the earlier Sherwin stage as well.

The name Aeolian Buttes till is proposed for this deposit. It is definitely pre-Bishop tuff at the Aeolian Buttes. I once thought it equivalent in age to the Sherwin till (Putnam, 1938, p. 82), but after visiting the type locality at Sherwin Hill 30 miles southeast (Blackwelder, 1931, p. 895–900), I believe, as Blackwelder does that the Sherwin till overlies the Bishop tuff.

The Aeolian Buttes till may well be equivalent in age to the McGee stage (Blackwelder, 1931, p. 902–906), but I prefer to use the proposed name because of the uncertainty of correlating between the two deposits and because of the better stratigraphic evidence here. The Aeolian Buttes till can be dated with respect to both the local volcanic and glacial succession; the McGee till at the type locality 25 miles southeast—scattered, deeply-weathered granitic erratics on a high ridge of hornfels—cannot.

Bishop tuff.—Bishop tuff (Gilbert, 1938) is a wide-spread volcanic rock underlying the area mapped east of the Sierra Nevada almost as far north as Mono Lake. This welded tuff has strikingly similar characteristics over a wide area. According to Gilbert, the tuff extends 400–450 square miles and has a total volume of approximately 35 cubic miles.

Bishop tuff is exposed near Aeolian Buttes and between the Mono and Inyo craters. Its exposure in the Mono Craters tunnel (Fig. 6) shows that it is continuous between these areas and extends under the Mono Craters. Smaller outcrops project through Tahoe and Tioga till along the Sierran margin, and isolated outcrops, possibly fault slivers, are found well up on the Sierran escarpment.

The tuff characteristically weathers pink. In forested sections, where podsolization is probably active, it is grayish-white and difficult to tell at a distance from granitic rocks. Below the weathered zone, the tuff is gray to black.

The matrix is consolidated rhyolitic ash. Embedded in this dusty groundmass are pumice fragments, some as much as 1 foot wide; perhaps 1–3 inches is average. According to Gilbert (1938, p. 1835):

> "Pumice fragments which are unoriented and very porous at the top of the tuff are progressively collapsed, flattened, and alined in the horizontal plane toward its base. At the bottom of thick sections, the fragments are lenticular; some are glassy and resemble lenses of porphyritic, perlitic obsidian, while others are lithoidal and spherulitic. The phenocrysts in these pumice fragments show no change."

The flattening of pumice fragments is a striking phenomenon where the tuff is exposed in the Mono Craters tunnel. The fragments are collapsed, have a horizontal orientation, are jet black, and resemble obsidian containing phenocrysts of quartz and sanidine.

The tuff also contains fragments of bedrocks, for the most part chips of hornfels, quartz monzonite, and some volcanic rocks, chiefly basalt. None are large, 4 inches appeared to be maximum, and all could have been derived locally. The tuff is cut by prominent joints which intersect in a pattern resembling closely fitted masonry.

Glaciated tuff has a smoothed surface as a rule. Joints are accentuated by ice erosion, so that blocks separated by them became rounded and may resemble granitic erratics or boulders of weathering. Ice tended to strip away the tuff in benches; this was controlled by depositional differences, or more likely by plucking along horizontal joints.

Nonglaciated tuff commonly shows cavernous weathering especially on the lee or shady side of rocks. Wind-fluting and pitting are well developed in the tuff on the west side of the Aeolian Buttes because this site was fully exposed to the periglacial winds of Rush Creek and Reversed Creek glaciers (Pl. 4, fig. 3). Cavernous weathering is also characteristic of outcrops at less exposed sites and is especially common along joints. Collapsed pumice fragments, usually less resistant than the matrix, weather to give the rock a Swiss-cheese appearance; uncollapsed shards, locally more resistant than the matrix, give the rock a studded surface.

Sherwin till.—Only two small areas of Sherwin till are in the area mapped (Pl. 1, 3); one is on the high ridge west of Rush Creek narrows, the other southwest of Williams Butte. The Sherwin till has been so deeply eroded that morainal forms are largely destroyed and only an undulating surface survives over weathered boulder clay.

At the type locality, about 30 miles southeast at Sherwin Hill (Mt. Goddard quadrangle), Sherwin till rests on Bishop tuff (Blackwelder, 1931, p. 899). In that area, as well as in the Mono region, the till is weathered and its original constructional form destroyed.

The till west of Williams Butte is marginal to a more extensive body of Sherwin till that covers most of the ridge between Walker and Leevining creeks. The till makes undulating sage-covered slopes that largely conceal Sierran bedrock (Pl. 7, fig. 1), except where some of the many faults that cut the till have large enough displacements to bring the underlying bedrock in view, or where stream deposits cut through the glacial debris.

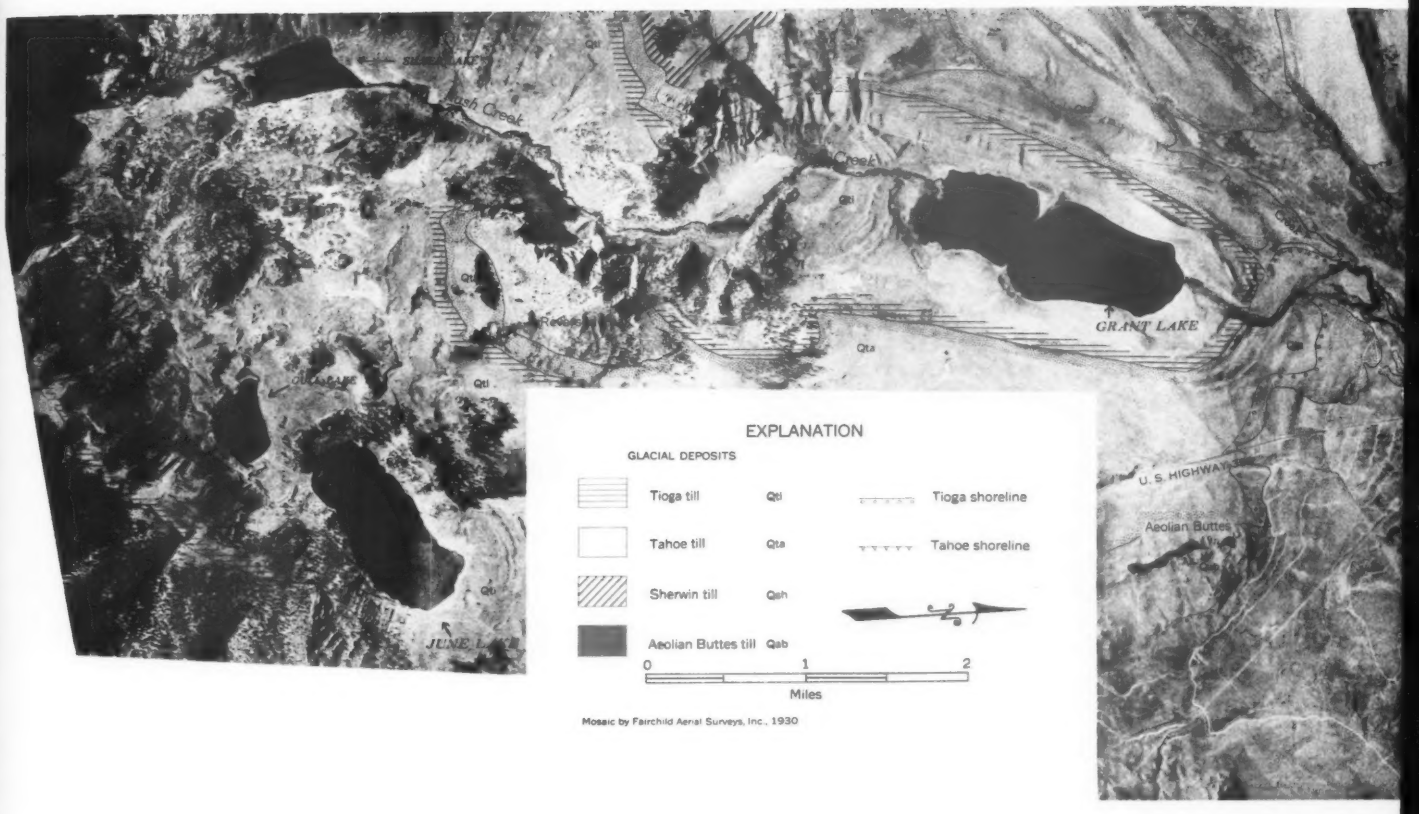
The second locality is west of Rush Creek narrows on a prominent bench at an altitude of 9100 feet. The till there is farther upslope than the limit of Tahoe glaciation and consists of deeply weathered boulders embedded in a clayey or silty matrix.

The Sherwin till according to Blackwelder (1931, p. 918) is equivalent to the Kansan stage in the central United States. Considering its great bulk, as well as the fact that more than one advance is represented, it may be correlative with the Illinoian stage as well (Blackwelder, 1931, p. 918).

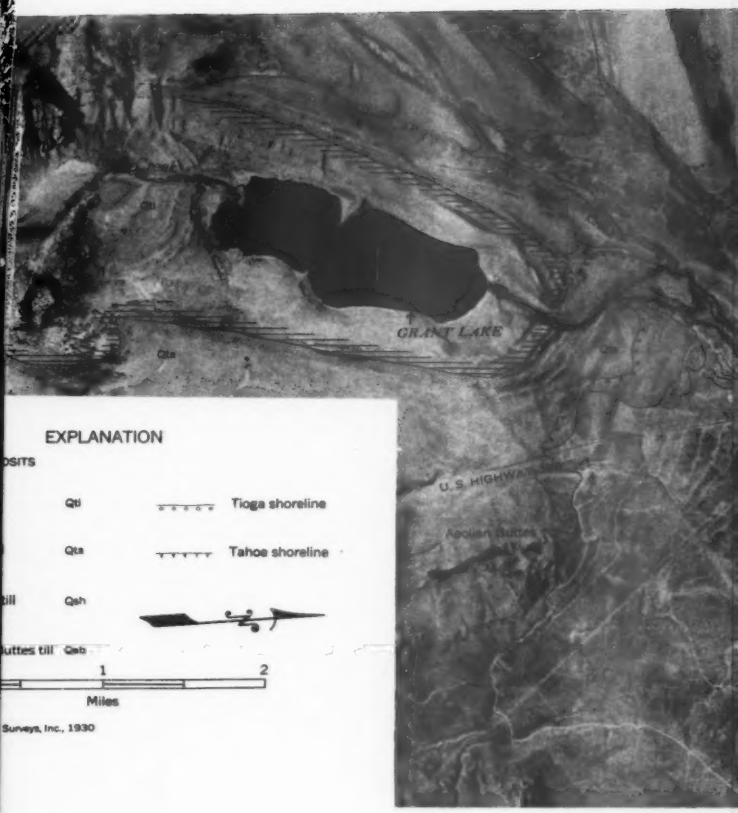
Tahoe till.—The bulky moraines of this stage are an impressive sight along the Sierran front. Their sage-covered slopes make a nearly continuous piedmont apron from June Lake to Walker Creek. These are the oldest glacial deposits in the area preserving their original morainal forms. Only lateral moraines survive; end moraines are eroded away. The lateral moraines are very large; for example, those bordering Grant Lake have a maximum height above the lake of 850 feet, are 3 miles long, 2500 feet broad at their base, and their crests have a longitudinal gradient of about 6 per cent.

Tahoe till is slightly more weathered than younger Tioga till, but does not have a mature soil profile. The top 2–10 feet of till may be stained tan or yellow-brown, as compared to the gray-blue or ash-gray Tioga till. Tahoe till consists of angular to rounded boulders in a stony-silt matrix. Many of the boulders are rounded and much of the till appears to be water-laid. Large erratics are relatively rare in the till. They appear more common than is actually the case because of their concentration on the surface of the moraine through removal of the finer matrix.

Bulky Tahoe stage moraines make ridges that extend each of the principal canyons beyond the Sierran front. All enclose lakes, except the



GEOMORPHIC MAP OF THE JUNE LAKE DISTRICT, CA



GEOMORPHIC MAP OF THE JUNE LAKE DISTRICT, CAL

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Tahoe moraines at Leevining Canyon where the morainal ridge is lacking on the north side and only a plaster of boulders survives on the rock wall of the canyon. The south-side moraine is one of the large ones in the area and is cut on its outer slope by a conspicuous set of Lake Russell shore lines.

Rush Creek glacier divided at Reversed Peak into (Fig. 5) Grant Lake arm and June Lake arm, and the June Lake arm forked again at the Aeolian Buttes. Grant Lake arm and the western branch of June Lake arm overlapped, spread as a thin sheet over the western flank of the Aeolian Buttes and extended into Pumice Valley as a piedmont glacier. The eastern half of June Lake arm advanced as a narrow prong into the area later occupied by the Mono Craters. In fact, as Russell pointed out (1889, p. 376), one of the larger obsidian flows crosses the lateral moraine of this arm (Pl. 5, fig. 2) and extends into the trough once occupied by ice.

An unusual feature of the Tahoe moraines at Walker Creek is an older set of lateral moraines that look as though they were an offshoot of the moraine (Pl. 7, fig. 1) bordering the eastern side of the valley. These subsidiary ridges may be moraines deposited during an earlier advance of the ice, followed by a retreat and a later advance on not quite the same course. The result is that the younger moraines override the older. The age of this earlier set of moraines is uncertain; too much of their constructional form seems preserved for them to be Sherwin. They are more likely a sub-stage of the Tahoe.

Basalt.—Three small basalt flows are in the area; the first near the southern margin between the Inyo Craters and the Sierran front, the second east of June Lake in the trough of June Lake arm of Rush Creek glacier, the third in Pumice Valley north of West Portal. The June Lake flow rests on Tahoe till and is covered for half its length by Tioga till. Most of the first flow is buried by the southern of the two Inyo obsidian domes. Basalt fragments scattered as ejecta in pumice associated with the dome demonstrate that this dome erupted through the basalt.

The Pumice Valley basalt is older than the other flows. Although most of the flow is buried by Mono Craters pumice, it appears to extend under the andesite dome at the north end of the chain. This field relationship is borne out by contamination of the andesite with augite and labradorite probably picked up from underlying basalt.

Some of the basalt is scoriaceous, much is vesicular. The vesicular basalt is commonly dark red; denser facies can be either dark red or black. Some basalt is porphyritic and the visible phenocrysts are chiefly feldspar laths.

The flow at June Lake is the most significant historically. Its source is a breached cinder cone that rises about 300 feet above the lake. The flow has issued from a breach in the north crater wall and, like Rush Creek glacier, it divides at the Aeolian Buttes. The western arm is the shorter; the eastern arm reaches and extends under obsidian flows of the Mono Craters.

Spatter cones are characteristic of the unglaciated part of the flow. Two particularly striking ones are on a north-south ridge that crosses the mid-part of the flow. This ridge may be due to a buried Tahoe recessional moraine, but more likely to a temporary blocking of the flow against a post-Tahoe fault scarp. The spatter cones were once more common; most of them have been quarried for road metal and decorative rock. Russell (1889, p. 345) believed each of the cones was the site of a separate eruption, but their origin as features on the surface of a flow was recognized by E. B. Mayo (Blackwelder, 1931, p. 891).

The till-covered cinder cone with a few striated ridges of rock near its rim is the only outcrop of basalt in the glaciated half of the flow (Pl. 5, fig. 3). The large number of basalt fragments in the Tioga till is indirect evidence that basalt underlies the glacial debris. Drumlin-like mounds in the Tioga terminal moraine may be till-covered spatter cones.

Tioga till.—Tioga moraines are nested inside the much larger Tahoe lateral moraines. The Tioga moraines consist of a family of terminal moraines and individual or groups of recessional moraines inside the glacial trough. Lateral moraines paired with these commonly extend as sloping ridges up the inner face of the much larger Tahoe laterals. Tioga terminal moraines are low; the average height above the valley floor of Leevining Canyon is 90 feet, and at Grant Lake about 100 feet. Recessional mo-

raines are lower; 85 feet is the maximum above Grant Lake, 60 feet is the maximum for Leevining Canyon. Recessional moraines are irregularly spaced and commonly clustered in groups. Examples are a morainal swarm where Grant Lake is nearly divided into two parts and the low ridges between June Lake and the basalt cinder cone.

There is little to distinguish between Tahoe and Tioga till in road cuts. The surface zone of tan or light-brown stained rocks is commonly a little shallower in Tioga till—perhaps 1–5 feet. The depth of pumice mantle is not a certain criterion either (Blackwelder, 1931, p. 891). On Tioga moraines adjacent to the Mono Craters it may be 10–15 feet deep; on more distant Tahoe moraines, about 1–2 feet. Depth of burial depends more on distance from the craters and topography than on age. Pumice is thinner on morainal crests, deeper in intermorainal depressions.

The best exposure of Tioga till is the large cut at the west abutment of Grant Lake dam (Pl. 7, fig. 2). Perhaps the most striking feature of this cross-section of a terminal moraine is the small number of large erratics; only one has a volume roughly comparable to an automobile (Pl. 7, fig. 2). The erratics are more commonly rounded than angular and only a few have striated surfaces. Many layers of till appear to be water-laid and on the lake side of the moraine the initial dip is west or towards the glacier front.

The matrix consists chiefly of silt and sand. The grains in these finer sediments are strongly angular and when slightly compacted are well stabilized. Aqueduct tunnels driven through the Grant Lake moraine required little timbering and stood several years before they were concreted. Matrix of till that was part of the ground moraine is locally laminated and has a hackly surface. This fissility may result from pressure of the over-riding ice.

Terminal moraines make low but conspicuously looped ridges inside the Tahoe laterals. There are 32 ridges at Leevining and 27 at Grant Lake. At June Lake the situation is complicated because the cinder cone interfered in the path of the glacier and also because the glacier expanded into the open trough where the Tahoe ice divided at the Aeolian Buttes. There appear to be 25 ridges, but the actual number is difficult to determine because of the thick pumice mantle.

All the moraines are breached by active streams, except at June Lake. Its former outlet can be traced through a low wind gap west of the basalt cone and then by a dry canyon now followed by U. S. Route 395. Channels across other terminal moraines are Rush, Walker, and Leevining creeks. Parker Creek is unusual because its channel crosses the Tahoe lateral moraine rather than the Tioga terminal. The low point in the morainal barrier is at the north abutment of the terminal moraine and here the stream has cut its spillway across the Tahoe lateral, leaving the terminal moraine intact.

Mono-Inyo craters obsidian.—The Mono Craters are an eastwardly convex chain of rhyolitic obsidian domes and flows that extend from Mono Lake southward 10 miles to Route 395. There are about 20 domes; the highest has an altitude of 9164 feet and rises about 2200 feet above Pumice Valley. The Inyo Craters, starting at U. S. Route 395, extend for 10 miles south. Only the northern two volcanoes of this chain are within the area mapped.

The obsidian shows a wide variety of types. The familiar jet-black glassy kind is relatively rare and commonly limited to minor flows or to individual protrusions on some of the domes. Most of the obsidian is pumiceous and consists of a fine network of glass fibers with a texture much like that of glass wool. There is a wide color range: light silvery gray, dark gray, purplish gray, purplish brown, and brownish red, with light grays and browns dominant. Some of the frothy obsidian consists of nothing but finely-spun glass; other varieties are abundantly porphyritic, and phenocrysts of sanidine, hornblende, and biotite are common. Microcrystalline obsidian, a darker red-brown or purple-brown rock, crops out in the throats of explosion craters that have destroyed some of the earlier domes.

A minor feature of interest is the presence of rounded cobbles and gravel on some of the domes and flows. These accidental ejecta are most common at the northern and southern ends of the Mono Craters. Those at the northern end are probably from sediments of Lake Russell and their origin was recognized by Joseph

FIGURE 1. ROCHE MOUTONÉE OF QUARTZ MONZONITE WITH CHARACTERISTIC SHEET JOINTING
Foreground slope covered with Aeolian Buttes till.

FIGURE 2. TYPE LOCALITY OF AEOLIAN BUTTES TILL
Boulders in foreground are till resting on smoothed surface of quartz monzonite. Rocky escarpment is Bishop tuff resting on till.

FIGURE 3. OUTCROP OF BISHOP TUFF
Showing cavernous weathering and wind-fluting.

FIGURE 1. NORTHERNMOST OF INYO CRATERS
Characteristic form of obsidian plug dome.

FIGURE 2. OBSIDIAN FLOW IN CENTRAL PART OF MONO CRATERS
Low ridge in middle distance crossed by flow is Tahoe lateral moraine of June Lake arm of Rush Creek glacier.

FIGURE 3. TIOGA ERRATIC RESTING ON STRIATED SURFACE OF INTERGLACIAL BASALT FLOW AT JUNE LAKE

Le Conte (1879). Cobbles on the summits of the southern craters are probably from stream gravel and outwash associated with the Aeolian Buttes till, as these deposits are encountered in the Mono Craters tunnel 1500 feet below. Bedrock fragments, chiefly quartz monzonite, are scattered on the surface of the southern Mono Craters, and, as mentioned before, basalt fragments are common in pumice associated with the southern of the two Inyo Craters.

The land forms of these rhyolitic volcanoes are described elsewhere (Russell, 1889; Williams, 1932; Mayo *et. al*, 1936; and Putnam, 1938). The basic form of this stiff, blocky lava is the plug dome, an unsatisfactory term for these cidadel-like structures (Pl. 5, fig. 1) with their near-vertical walls girdled with a continuous mantle of blocky talus and surmounted by a rampart of obsidian pinnacles. They look more like a truncated cone.

The rise of one of these so-called domes toward the surface probably is heralded by gas explosions, capable not only of excavating explosion pits but of hurling vast quantities of pumice into the air. Commonly an explosion pit is surrounded by a tuff ring, highest on the eastern or leeward side.

Every gradation exists from explosion pits with no visible obsidian in their center to craters that have been completely buried by the growth of an obsidian plug. An excellent example of an explosion pit, this one blasted through Bishop tuff, is at the foot of Deer Mountain, midway between the Mono and Inyo craters. The pit is 200 feet deep and about 1500 feet across. Panum Crater, nearest Mono Lake, is an example of an obsidian plug that almost fills the explosion pit and yet is surrounded by an unbroken tuff ring.

The two Inyo "Craters" are obsidian domes and both lack an encircling lapilli rim. The smaller one, next to Highway 395, is about 500 feet high and 2400 feet through at the base. It consists of gray, pumiceous and blocky obsidian. The southern dome is about 700 feet high and 5000 feet in diameter and has an extraordinarily jagged surface of glassy and streakily flowbanded obsidian.

Some obsidian protrusions have expanded beyond the original explosion pit as blocky steep-walled flows, leaving the lapilli rim complete except at the breach. Two of these flows are outstanding: one, near the northern end of the Mono Craters, is 2 miles long and 400 feet high at the toe; the second, near the center of the chain, started from a fissure at the crest and has flowed east and west from this source.

The mechanics of advance of one of these flows is well described by Russell (1889, p. 383):

> "The extreme ruggedness of the coulees [flows] is due to the fact that they hardened at the surface during the time they were still moving. The crust thus formed became broken and involved in the pasty material beneath in the most complicated manner. The steepness of the scarps formed at the ends and sides of the coulees was also due to the viscid condition of the glass composing them. In flowing down the side of the craters the lava descended slopes that must have had an inclination of fifteen or twenty degrees, but only in the case of the greatest eruptions did the viscid streams reach the plain at the foot of the cones. In no instance did they continue their course for a considerable distance after leaving the abrupt slopes. . . ."

Explosions were not limited to the preliminary phases of an eruption. An excellent example of an explosion that nearly demolished an earlier obsidian dome is the large crater immediately south of the Mono Craters tunnel. This explosion crater, interpreted by Russell (1889, p. 382) as the result of collapse, clearly reveals the internal structure of an obsidian dome. It can be seen that the obsidian rose through a narrow vent and expanded outwards at the surface to give a structure roughly comparable to a sheaf of grain.

Other post-dome explosion pits are near the crest of the range. The most notable examples are (1) immediately south of the two conspicuous central domes and (2) at the head of the large northern lava flow. Pumice from these pits has been showered deeply over domes and craters, most conspicuously at the northern lava flow whose western end is buried by a great ash slide, 1200 feet high and with a nearly continuous slope of between 30° and 35°.

That some of the Inyo and Mono craters are post-glacial is indicated by two facts: (1) the northern craters do not have shore lines of Lake Russell cut on their slopes, although they are below the level of even the lower strand lines, and (2) pumice from both the Inyo and Mono craters covers late-recessional Tioga moraines.

Pumice and alluvium.—Pumice from the Mono and Inyo craters blankets much of the

area. Depth depends on proximity to the craters, topographic setting, and relation to the prevailing wind. Pumice is 20 feet thick in Shaft 1 of the Mono Craters tunnel about 0.5 mile east of the craters, on nearly level ground to the lee of the volcanoes. On the windward side at a comparable distance, pumice is seldom more than 5 feet thick.

Pumice is a real obstacle to geologic mapping in the vicinity of June Lake. As a result all contacts here must be regarded as approximate, and it is a hindrance as far north as the Aeolian Buttes. Pumice is practically absent north of Rush Creek on steep slopes, although it covers low ground along Mono Lake. In contrast to its rather meager distribution west of the craters, wind-blown pumice was drifted east as far as the Nevada boundary, 30 miles from the source.

Included in the pumice are obsidian fragments that range from dust to blocks; lapilli are extremely common. Particle size is roughly proportional to distance from the source, but density is also important. Blocks of black, glassy obsidian up to 6 inches wide are common near the craters; with increasing distance obsidian fragments are rarer and most of the pumice is the familiar gray, frothy variety.

A unique feature of the pumice is its resistance to erosion, probably because of the naturally high permeability not yet reduced by the development of a soil profile. Although the pumice is several thousand years old, there are wide areas, including much steep ground, where no stream channels have been cut.

Alluvium is limited to stream gravel, chiefly in the larger creeks, and to talus cones at the base of high cliffs, conspicuous in Rush Creek narrows and at the toe of the June-Grant horseshoe. Alluvium in Rush and Reversed Creek valleys includes some till in the form of ground and recessional moraines. There are several low recessional moraines in Rush Creek valley between the narrows and Silver Lake.

Structure

Faults are the most important Quaternary structural features. Faults cutting glacial till are commonly more apparent with a stereoscope than on the ground. They are identified by low scarps, and in places by linear depressions, by springs, and by abrupt changes in vegetation.

There is no single Sierran fault in the June Lake district; instead faults make a rectilinear pattern with one system striking north-south to N 20° W, and the other striking from N 30° W to N 60° W. The lack of any single large-scale fault is confirmed by the map pattern of the range front: there is no simple linear escarpment—it consists of bold salients and deep reentrants. This is especially well-shown at Williams Butte where the northwest-trending Leevining escarpment ends abruptly and the range front retreats 5 miles westward to Walker Lake where it once more resumes the interrupted trend.

Nor is the eastern Sierra a continuous escarpment from base to summit, but in profile rises as a number of great benches separated by steeper slopes. Russell (1889, p. 277) noted this step-like appearance, most striking west of Leevining where the steep escarpment at the western shore of Mono Lake makes the riser below the broad tread of the bench on the east slope of Mt. Warren at about 10,000 feet.

Displacement on any individual fault is small, but the cumulative effect is great. Most of the faults shown on the map probably have a throw of only a few hundred feet. The Sierra Nevada has not been elevated here along a single fracture but has been uplifted by repeated slight movements on a host of minor, closely spaced faults.

Faults tend to feather outwards from a zone where they are closely spaced to one where they are widely-separated. Faults of this type probably are distributive in that the throw decreases away from the zone of close-spacing.

Movement on many of the faults has been recurrent, tending to decreased vertical displacement with recency in time. This is demonstrated where they cut moraines. The throw on a given fault can be 100 feet or more where Sherwin till is offset, as near Leevining Canyon, perhaps 20 feet in Tahoe till, and none, or only a very slight amount, where the fault is covered by Tioga till.

Much of the displacement along the Sierran front occurred before the Sherwin glacial stage (Gilbert, 1938, p. 1832). The Sierran escarpment is highest west of the shoulder blanketed by

Sherwin till west of Williams Butte. Both Sherwin till and Bishop tuff are faulted, but the amount of displacement is not as great as the earlier movement.

There are two zones of closely-spaced faults in the June Lake district: the outer along the eastern face of Williams Butte, Reversed Peak, and the granitic mountains southeast of June Lake; the inner along the east face of Mt. Parker, Mt. Wood and San Joaquin Mountain. The outer zone has been recently active because faults cut Tahoe and, indeed, some Tioga moraines. The inner fault zone strikes through Silver Lake, as described in a later section, and this has controlled the drainage reversal of June and Grant Lake arms of Rush Creek glacial trough by forming an easily eroded crush zone.

Where actual fault planes are exposed, as on walls of deep glacial canyons, the dip ranges from vertical to about 70°; an example is the prominent glacially scoured fault plane south of Silver Lake (Pl. 6, fig. 2). Although the camera axis is not in line with the plane of the fault, it is close enough to show that the dip is considerably greater than the slope of the unglaciated mountain front above the fault trace. Average slope of nonglaciated parts of the Sierran escarpment is more moderate than the dip of the faults; for example, the mountain front west of Leevining is about 26°–27°.

There are definite indications of faulting in the lower country east of the Sierra Nevada. The Inyo and Mono craters are almost certainly aligned along a fracture system. The Mono Craters tunnel penetrates a shatter zone directly beneath the axis of the craters. Explosion pits in the gap between the Inyo and Mono craters seem to be controlled by faults at the base of Deer Mountain. This mountain and the Aeolian Buttes are probably horsts, but their structure cannot now be determined because of pumice that mantles their slopes.

GEOMORPHOLOGY

Lake Russell

Mono Lake.—Mono Lake is a prominent feature in the present landscape, but its predecessor, Lake Russell, was even more impressive in Pleistocene time. The modern lake has an average area of 85 square miles, measuring 10 miles wide by 14 miles long. The maximum depth, according to Russell (1889, p. 287), is 152 feet between Paoha Island and the Mono Craters. The lake water is unpleasant to taste and smells foul. It contains nearly equal amounts of NaCl (18.51 gm/liter) and Na_2CO_3 (18.67 gm/liter) according to an analysis made by T. A. Chatard in 1888 (Russell, 1889, p. 293). The lake level has fluctuated in historic time and may be higher now than it was in Russell's day, for a search by engineers of the Los Angeles Department of Water and Power failed to discover the mark cut in rock at water level on the northern island in Mono Lake by W. D. Johnson (Russell, 1889, p. 299). The lake level is now falling and probably will continue to fall as water withdrawn from Leevining and Rush creeks is diverted into Mono Craters aqueduct.

Shore lines.—Former lake shore lines on the Sierran slope behind Leevining are obvious enough to impress the most casual observer. They rise to an approximate altitude of 7070 feet, about 654 feet above the surface of the lake (the altitude of Mono Lake was 6416 feet, August 20, 1947).

The faint trace of an older, higher shoreline whose altitude, less certainly established, appears to be between 7170 and 7180 feet is tentatively correlated with the Tahoe glacial stage and the lower with the Tioga.

Locations of surviving stretches of Tahoe and Tioga shore lines are shown on the geologic and geomorphic maps (Pl. 1, Pl. 2). The faint Tahoe bench is preserved at 5 localities: (1) north of Leevining as an ill-defined shoulder on the mountain front; (2) south of Leevining Creek as a moderately prominent bench on the Tahoe lateral moraine; (3) on the Parker Creek Tahoe lateral moraines; (4) at Rush Creek as a rather broad, gravel-covered bench on the northern Tahoe lateral moraine; and (5) at the Aeolian Buttes as a bench cut in Bishop tuff.

The altitude of the lower Tioga shore line is well-established where the former meltwater channel of the June Lake arm of Rush Creek glacier reached Lake Russell. At this point, U. S. Highway 395 crosses the Mono Craters aqueduct. U. S. Coast and Geodetic Survey Bench Mark W-123 (1933)—elevation 7071.5 feet—is 330 feet west of the highway here. The Tioga

shore line is probably within ±5 feet of the elevation of the bench mark.

The Tioga shore line not only cuts outer slopes of Tahoe moraines but extends inside the former glacial troughs. It is in contact with Tioga moraines at Rush Creek and Leevining Creek. At neither of these localities has a lake cliff been cut in the moraine. At each point of contact, meltwater streams built up a delta that merged with outer slopes of the Tioga terminal moraine within the bay formed by the Tahoe lateral moraines.

There are approximately 30 benches below the Tioga shore line. Some are scarcely more than faint strand lines, but others are broad platforms, such as the surface of the delta built by Rush and Leevining creeks. I hope to determine later the relation of these lake levels with terminal moraines of nearby Sierran glaciers. The most conspicuous of the lower strand lines lies between 6700–6800 feet, the altitude of most of the broad plain of Pumice Valley and of the larger deltas of Rush and Leevining creeks, including the bench on which the town of Leevining stands.

Whether or not Lake Russell had an outlet is an important problem. Russell did not make a direct statement, except to indicate the lack of connection with Lake Lahontan to the north. He did point out the low relief of the divide separating the southeastern part of the lake from Adobe Valley (1889, p. 300). To see if this were a possible outlet I made a trip to this drainage divide with Mr. Claude James, Hydrographer, Los Angeles Department of Water and Power.

The evidence shows that Lake Russell overflowed into Adobe Valley in the older Tahoe but not the younger Tioga stage. A channel is cut in basalt across the Mono Basin divide at an approximate altitude of 7150 feet east of the abandoned Benton-Bodie stage road (Sec. 3, T 2 N, R 29 E, Hawthorne Quadrangle). This spillway connects with Adobe Valley by a number of linking shallow basins and ultimately by a deep, narrow canyon also eroded in basalt (Pl. 3). A chain of lakes in the eastern part of Adobe Valley appears to be part of a once-integrated drainage system. A slot-like gorge connected with the lakes and cutting through the Benton Range west of Benton (Mt. Morrison quadrangle) gives Adobe Valley a direct link with the Owens Valley. Therefore, during the Tahoe glacial stage, Lake Russell was the northern end of a chain of lakes, channels and spillways that extended from Owens Valley via Searles and Panamint lakes to Lake Manly in Death Valley. Hubbs and Miller (1948, p. 78–79) and Miller, (1946, p. 43–47) have already shown the bearing of this integration on the problem of fresh-water fish distribution in the Great Basin.

Lake sediments.—Only an incomplete section of lake sediments is exposed in the area mapped. The best sections are exposed in the canyon walls of Leevining and Rush creeks; in Rush Creek, 160 feet of beds are exposed from the canyon floor to the 6800-foot terrace surface of its delta. This incomplete section does give a good idea of the type of sedimentation near the shore of Lake Russell close to the mouth of a large glacial stream.

Much of the sediment, perhaps 60 per cent, is silt; there is almost no clay, and the remainder of the material is sand and gravel. The silt is dominantly light gray and appears to be derived from granitic rocks. The sand is commonly iron-stained and is generally darker; the gravel, which consists chiefly of hornfels fragments, is dark gray or brown.

The lake sediments are noteworthy in their large number of gravel beds; there are five near the mouth of Rush Creek. Russell, doubtless influenced by the idea of two periods of expansion of Lake Bonneville separated by an interval of desiccation, believed that the single layer of gravel he found, apparently near the present site of Leevining, marked a dry interval separating two pluvial periods (1889, p. 306–307). The large number of gravel beds intercalated with lacustrine silt is less likely to be related to periods of desiccation than to times when the mouths of these creeks were close to the place where the gravel occurs. Antevs (1926, p. 99) believes that the alternation of coarse and fine sediment is primarily related to fluctuations in lake level, although he does recognize what seems to me the more likely interpretation, that changes in the position of stream mouths is responsible for the abrupt change from gravel to silt.

Strongly-contorted beds are abundant in the

FIGURE 1. GULL LAKE
View west towards source of Rush Creek glacier. Upper limit of glaciation is at transition from smooth to craggy slope.

FIGURE 2. LOOKING SOUTH ACROSS SILVER LAKE
Direction of ice advance from right to left. Upper limit of glacier is skyline notch on high peak. Slot-like valley beyond lake is fault line etched out by ice.

FIGURE 3. VIEW NORTH ACROSS SILVER LAKE AND DOWN VALLEY OF RUSH CREEK
Ice moved northward. Upper limit of glacier is rocky face of highest peak.

REVERSED CREEK — RUSH CREEK TROUGH

FIGURE 1. SIERRA NEVADA AT WALKER CREEK
Embankments at canyon mouth are Tahoe lateral moraines; Tioga terminal moraines are nested between them. Rounded slopes below snow and at right margin are Sherwin till. Foreground slopes are deposits of Lake Russell.

FIGURE 2. CROSS-SECTION OF TIOGA TERMINAL MORAINE AT GRANT LAKE DAM

FIGURE 3. CONTORTED LAKE BEDS NEAR MOUTH OF RUSH CREEK

GLACIAL AND LACUSTRINE DEPOSITS

lake sediments. These zones of crumpled sand and gravel are bounded above and below by undisturbed strata (Pl. 7, fig. 3.) Russell (1889, p. 308–310) discussed them at length and concluded that they had a variety of origins including ice-shove and slippage along bedding planes. Slippage seems the more likely origin, for they appear to result from outward-gliding of saturated deltaic deposits as the result of overloading. Almost all the folds are overturned towards the lake, away from the source of sediment and opposite to the direction probable if produced by ice-shove.

Origin of Reversed-Rush Creek Canyon

An anomalous topographic feature of the area is the horseshoe valley of Reversed and Rush creeks occupied in part by June, Gull, Silver, and Grant lakes. June Lake (7630 feet) is the highest of these; Grant (7116 feet, August, 1947[1]), the lowest. Reversed Creek is aptly named; it flows from Gull Lake towards the mountains instead of towards the low country east of the Sierra Nevada. Reversed Creek joins Rush Creek at the toe of the horseshoe and the combined streams continue on to Grant Lake in the western arm of the loop. June Lake has no effective outlet other than a small drainage canal cut across the marshy ground separating it from Gull Lake.

The origin of such a unique topographic feature naturally has invited speculation. Russell (1889, p. 343) believed that the reversal of flow in June Lake arm resulted from the damming of the stream by June Lake terminal moraines of the Tioga stage, and that a body of ice, trapped at the site of June Lake and isolated by the retreat of Rush Creek glacier, supplied the water necessary for Reversed Creek to cut the 400-foot step from Gull to Silver Lakes. Salisbury and Atwood (1908, p. 76), in a paper based on map interpretation, followed Russell but believed that reversal was accomplished by damming of runoff by the isolated ice body at June Lake.

Kesseli (1939) effectively disposes of these arguments by pointing out the unlikelihood of a small glacier at June Lake surviving at the same time that the parent Rush Creek glacier retreated into the high mountains west of the horseshoe valley. Had the isolated body of ice survived, it is even more unlikely that it could remove a mass of rock equal to about half the volume of the ice.

Kesseli proposed a new explanation, involving the following elements: (1) the horseshoe shape of the trough is inherited from an entrenched meander cut by a large river; (2) the original direction of flow of this river was from Grant Lake arm into June Lake arm (opposite to the present stream); (3) the meander was isolated from its parent valley by uplift along the marginal Sierran fault; (4) the present reversal of drainage results from tilting towards the northwest, probably as a result of this faulting; and (5) the difference between the steeper gradient of Reversed Creek (500 feet per mile) and the gentler Rush Creek gradient (5 feet per mile) is the result of stream rather than ice erosion. In short, the deepening of the western arm by 400 feet is the work of Reversed and Rush creeks.

Two main fault systems cross the area, the outer one along the Sierran front, the inner, and more important to the solution of this problem, along the escarpments of Mt. Parker, Mt. Wood and San Joaquin Mountain and under the trough occupied by Silver Lake. If we assume that two east-flowing streams crossed the area and were separated by a low divide at about the south end of present-day Silver Lake, the next step is that a subsequent tributary of one of these streams, by working headward in a crush zone along the inner fault system captured the other. The probability is that the western stream (Rush Creek) captured the eastern (Reversed Creek) because of the advantage of working in less resistant rock, chiefly shale hornfels. Whether or not the capture was completed before the Tahoe glacial stage must remain an open question. A nearly similar situation exists today on the divide between Glass and Deadman creeks south of June Lake where a subsequent tributary of Deadman Creek, working along the same fault zone that strikes through Silver Lake, has lowered the interstream divide and is on the verge of capturing the headwaters of Glass Creek.

[1] The altitude of Grant Lake has a considerable range since it is now used as a reservoir by the Los Angeles Department of Water and Power.

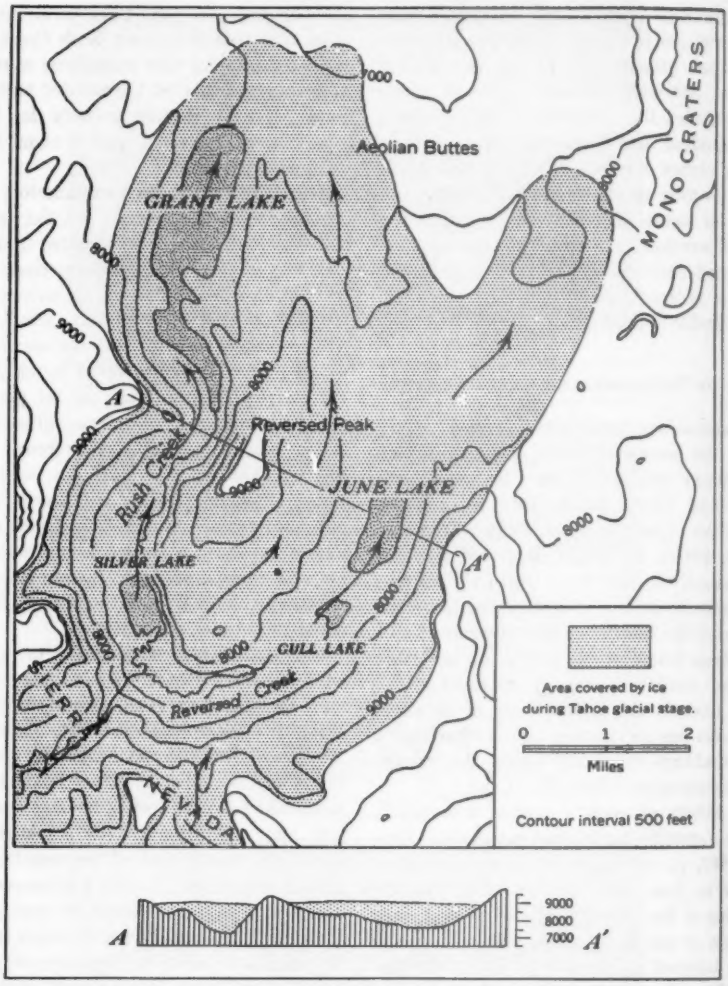

FIGURE 5.—RUSH CREEK GLACIER DURING THE TAHOE STAGE

The problem of the reversal of drainage is seemingly related to two factors: (1) the relative resistance of the bedrock, and (2) the relative erosional efficiency of the two branches of Rush Creek glacier.

Rocks in the June Lake drainage are relatively more resistant than those in the Grant Lake area. The geologic map (Pl. 1) shows a greater width of outcrop of quartz monzonite in the June Lake arm; this comparatively resistant rock crops out both at June Lake and at the west end of the glacial trough; but is exposed only at the narrows in Grant Lake arm. Also, in the vicinity of Grant Lake these rocks are chiefly shale hornfels with a near-vertical dip. Lime-silicate rocks and the massive, banded variety of hornfels have an outcrop width of only a few hundred feet west of the narrows. In June Lake arm, shale hornfels, a rock that is very susceptible to glacial erosion, has been almost cut out by the quartz monzonite intrusion whose contact cuts obliquely

across the strike of the shale hornfels. Dominant types of metamorphic rocks cropping out in June Lake arm are lime-silicate hornfels and massive, banded varieties derived from sandstone and limy shale; these rocks make rounded bosses and *roches moutonées* and are almost as resistant as quartz monzonite. There can be little question that June Lake arm of Rush Creek glacier was less effective in erosion.

The other major consideration (2) is the relative efficiency of the two ice tongues. Figure 5 shows that the June Lake glacial tongue averaged 2.2 miles wide, compared with 1 mile for the Grant Lake ice. The cross-section shows that not only was the Grant Lake arm deeper, 1800 feet compared with 1300 feet, but it was concentrated in a narrow, well defined channel. June Lake glacial arm used up much of its energy on the broad upland shoulder of Reversed Peak, and with the reduced volume of ice available in Reversed Creek had to attempt the reduction of much more resistant rock than in Rush Creek Canyon.

Rush Creek glacier had a comparatively easy task in Grant Lake arm and a more difficult one in June Lake arm. Depth of the glacial trough bears a close relationship to the nature and fracture pattern of the rocks; the trough is deeper at Silver Lake in the zone of most intensive faulting, shallower where the ice was forced over massive rock cut by widely-spaced joints immediately west of Gull Lake.

That the valley floor in June Lake arm slopes opposite to the direction of ice advance is not surprising. Flow of glacial ice is controlled by the slope of the upper surface, and there is abundant evidence in the lateral moraines bordering Grant and June lakes that this was northward. The bedrock profile beneath Yosemite Valley is an especially impressive example of ice that eroded uphill (Gutenberg and Buwalda, 1939).

Engineering Geology

Mono Craters Tunnel

The Mono Craters tunnel provided an unusual opportunity to obtain a cross-section of the geology beneath the Mono Craters. This 11.5 mile bore was driven by the Los Angeles Department of Water and Power as part of a conduit to take water from the drainage basin of Mono Lake to the Owens River and thence to Los Angeles. Work on the tunnel started in December, 1934 and the section between West Portal and Shaft 1 was holed through in August, 1938. The length of time required to complete this 5-mile section gives a hint of the great engineering difficulties encountered.

Rocks penetrated in this stretch are shown on the cross-section (Fig. 6). A ventilating shaft was sunk at Drill Hole 3J when the CO_2 concentration in the tunnel became dangerously high.

The first 5000 feet of the tunnel are in Bishop tuff and quartz monzonite; little difficulty was encountered in this section, except for a small buried valley filled with Aeolian Buttes till. For the next 9500 feet the tunnel is in Bishop tuff, except for 1200 feet in pumice overlying the tuff. A record of 90 feet of advance per day was set in this dry, unconsolidated material.

The Bishop tuff drilled readily, broke easily, and, where not too closely jointed, required little support. Yet the long section of tunnel in this rock proved extraordinarily difficult to excavate. Where highly fractured, the tuff contained a great volume of water. As much as 7000 gallons per minute flowed into the tunnel, and since the gradient is in the direction of advance all this water had to be pumped back to West Portal. The temperature of the water increased from about 35° F. where it was first encountered at the western edge of the Mono Craters to about 90°F. at their axis.

Carbon dioxide entering the tunnel was one of the most vexing problems. Its distribution was essentially the same as the section beneath the craters, about 12,000 lineal feet. The inflow of around 200 cubic feet per minute necessitated the sinking of an auxiliary ventilating shaft and the installation of a blower at its collar as well as at West Portal.

At a point almost directly beneath the axis of the craters, the tunnel passed through the vent tuff associated with an obsidian plug immediately north of the tunnel line. Here, the heading collapsed and a mass of ash, brecciated obsidian, and shattered tuff broke into the tunnel together with a volume of water that

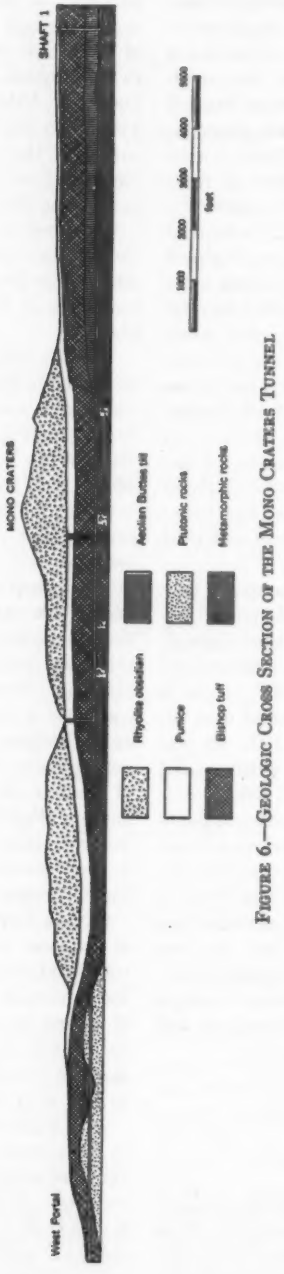

FIGURE 6.—GEOLOGIC CROSS SECTION OF THE MONO CRATERS TUNNEL

backed up for 10,000 feet. A swarm of closely spaced obsidian dikes was encountered at the same place; few of these were more than 6 inches across.

Shaft 1 proved extremely difficult to sink. The Bishop tuff, here about 500 feet thick, made no difficulty, but waterlaid sand and gravel immediately beneath it did. These unconsolidated sediments, probably glacio-fluviatile deposits associated with the Aeolian Buttes till, were saturated with water which flooded the tunnel when the walls collapsed. A year was spent in pumping water out, in preventing the nearly continuous caving of the sides, and in concreting the shaft.

The tunnel grade between Shaft 1 and the Mono Craters is close to an erosion surface cut across hornfels and later buried by about 350 feet of cobbles, gravel, and sand underlying the Bishop tuff. The sediments are saturated with water and completely unindurated. As a result the tunnel required continuous support at an average depth of 900 feet. Steel tunnel sets anchored in concrete were used and even though these were placed in contact with one another, failures were common. Headings in the incompetent material could be advanced only by breast-boarding and driving spiling ahead of the face.

Summary and Conclusions

Bedrock of the June Lake district consists of sandstone, shale, and limestone metamorphosed to hornfels by intruded quartz monzonite and related plutonic rocks.

These rocks are cut by high-angle normal faults. These faults are closely spaced and in many cases have undergone repeated movement along the same fracture. This movement has continued throughout the Pleistocene, and in some cases into the Recent, as shown by increasing throw where faults cut successively older moraines.

There is no single Sierran fault in the June Lake district. The rectilinear pattern of the range front is related to the fault pattern. The step-bench profile of the Sierran ridges is related to zones of relatively greater displacement within this fracture system. One of these fault zones strikes through Silver Lake trough; others are at the outer margin of the range at Reversed Peak, and at Leevining and Williams Butte.

Volcanic activity has been intermittent throughout the Quaternary and has alternated with glacial stages. The oldest volcanic rock seems to be biotite dacite on the ridge south of June Lake. This flow is cut by the Silver Lake fault system, here with a throw of several hundred feet.

The Aeolian Buttes till is the oldest glacial deposit and may be equivalent to Blackwelder's McGee stage. It crops out beneath the Bishop tuff on the Aeolian Buttes and Deer Mountain, as well as in the Mono Craters tunnel.

Bishop tuff, a rhyolitic ash and pumice welded to make a compact and homogeneous rock, covers about 400 square miles of country east of the central Sierra Nevada and was erupted after the deposition of the Aeolian Buttes till. Sherwin till, deposited following eruption of the Bishop tuff, covers some of the upland benches, notably between Leevining and Walker canyons. The morainal form is destroyed and only weathered glacial debris remains.

The last two glacial stages, the older Tahoe and the younger Tioga, are recent enough so that moraines survive. The Tahoe is characterized by bulky lateral moraines, the Tioga by moraines that are nested inside the Tahoe lateral embankments and also include fresh-looking terminal, lateral, and recessional moraines.

Mono Lake is a relic of a much larger lake during the Tahoe and Tioga glacial stages. This Pleistocene lake is here named Lake Russell for I. C. Russell as recognition of his pioneer work on the geology of the region. Two high-level shore lines can be recognized; the less well preserved at 7180–7190 feet is correlated provisionally with the Tahoe stage, the lower, 7070-foot level with the Tioga. During the Tahoe stage, but not the Tioga, the lake had an outlet at the southeastern shore and was connected with the Owens Valley-Death Valley drainage system.

Late in the Pleistocene and continuing in Recent times, the Mono and Inyo craters erupted large quantities of rhyolitic pumice and

obsidian. The volcanoes include explosion pits, craggy plug domes, and stubby, steep-sided flows. The Mono Craters are a compact group of superimposed flows and domes that rise 2000 feet above the surrounding plain. The Inyo Craters are separate domes that in some respects appear more recent than the Mono Craters. The absence of Lake Russell shore lines on the slopes of the northern Mono Craters and the presence of pumice on late recessional Tioga moraines from both chains of craters proves the recency of their eruption.

The unusual horseshoe-shaped trough of Reversed and Rush creeks seems more likely to be the result of pre-glacial stream capture and differential erosion by ice, than isolation of a large meander by faulting and its subsequent minor sculpture by glaciation.

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Short Notes

A reversed seismic refraction measurement was made in 2800 fms. of water at Latitude 34°N., Longitude 66°30′W. (120 miles northwest of Bermuda and 400 miles east of Cape Hatteras). The ocean floor was found to be composed of two layers. A velocity of 24,800 ft. per sec. (7.58 km. per sec.) was found for the second layer. This was identified with the P_n waves of earthquake seismology attributed to the basaltic layer found below the Mohorovicic discontinuity. A first layer was clearly present, but the velocity was not determined. A velocity of 5600 ft. per sec. (1.70 km. per sec.) was assumed, giving a thickness of 4500 feet for this sedimentary layer. This velocity was chosen from some of our earlier measurements, as yet mostly unpublished. The granitic and intermediate layers are thus absent beneath the ocean floor at this point. The velocity of the basaltic layer compares favorably with those of Slichter in New England and with Jeffreys' world average. A detailed report of this work is being submitted for publication to the Journal of the Seismological Society entitled, *Seismic refraction measurements in the Atlantic Ocean Basin, Part I*, by Maurice Ewing, J. L. Worzel, J. B. Hersey, Frank Press, and G. R. Hamilton.

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